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ANNOUNCEMENT

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JANUARY 1957

ECOLOGICAL INTERPRETATIONS OF PLIOCENE AND PLEISTOCENE STRATIGRAPHY IN THE GREAT PLAINS REGION

JOHN C. FRYE and A. BYRON LEONARD

ABSTRACT. In the Great Plains the succession of gross ecological conditions through Neogene and Quaternary time may be interpreted from the stratigraphy, geomorphological history, buried soils, and fossil mollusks and plants. A mild humid climate prevailed over a late-mature erosional topography in earliest Neogene. Progressive reduction of topographic relief and lowering of the regional water table, accompanied by an almost uniform drying of the climate, followed. Semiarid, rigorous conditions existed on the constructional plain, temporarily at erosional equilibrium, at the end of the Tertiary. A sharp reversal of climatic trend accompanied by stream incision and minor alluviation marked the beginning of the Pleistocene. The trend toward humidity culminating in the Kansan, was followed by a return to the trend—pulsating and irregular but none the less distinct—of progressive desiccation. The Recent climate is judged to approach in dryness and rigor that which existed on the late Tertiary plain.

INTRODUCTION

Paleoecology of marine sediments has received increasing attention by geologists. The paleoecology of terrestrial "flatlands" deposits, however, has been largely neglected in spite of the fact that the late Cenozoic strata of the continental interior contain a wide variety of interpretable data. From our studies of the stratigraphy and paleontology of the Neogene and Quaternary of the central and southern Great Plains have emerged data including an almost complete succession of fossil molluscan faunas, abundant plant remains through middle and late Neogene time, well preserved and extensive buried soil profiles from latest Neogene and Quaternary time, a distinctive sequence of clastic sediments, and the history of geomorphic development. It is our purpose here to reconstruct the succession of gross environmental conditions that obtained in this region through Pliocene and Pleistocene time.

The portion of this work dealing with Texas was supported by the Bureau of Economic Geology of The University of Texas as part of a continuing program on the Cenozoic geology of the northwestern part of the State.

The region considered extends from central-western Texas northward across western Oklahoma, Kansas, and Nebraska to the southern limit of South Dakota—a north-south strip approximately 200 miles wide and 700 miles long that includes much of the High Plains section of the Great Plains physiographic province. Its present aspect is a relatively undissected plateau, underlain by late Cenozoic deposits and segmented by the valleys of the through-flowing Canadian, Cimarron, Arkansas, and Platte rivers. The broad upland areas have a very coarse-textured drainage pattern and locally lack external drainage.

The climate is semi arid and windy; rates of evaporation greatly exceed precipitation rates. Average annual rainfall is as low as 15 inches in the southern part and nowhere exceeds 25 inches. Temperatures fluctuate widely; extremes of 118° to 120°F and –27° to –47°F have been recorded. The average annual afternoon wind velocity is approximately 15 miles per hour, which contributes significantly to the high evaporation rate.

The Great Plains are virtually treeless, although the hackberry, western cottonwood, willow, and some woody shrubs are found sparingly along water courses. The uplands support a cover of grasses among which buffalo grass predominates. Cacti and yucca are common, particularly in the south; sagebrush is common along water courses and on dune areas but in the south is replaced by mesquite and other xerophytic shrubs. The native fauna consisted of great herds of American bison and antelope, together with the prairie wolf, coyote, badger, prairie dog, and a great variety of small rodents that still inhabit the region. Molluscan populations are sparse and composed of few species; branchiate snails live in a few isolated localities where cold waters are perennially available from artesian springs, whereas the few widely distributed aquatic pulmonate snails are capable of surviving periods of drought. The grassy plains support few terrestrial snails, although a variety of pulmonate gastropods occur along stream courses where shrubs, mixed grasses, and herbs form the vegetative cover.

STRATIGRAPHY

The succession of environmental conditions is deduced from the deposits and their contained organic remains. The descriptive stratigraphy has been documented, but a statement of stratigraphic framework and a brief summary of depositional history and lithologies is requisite to ecological interpretations. Nomenclature differs somewhat from state to state but correlations within the late Cenozoic are now reasonably well established throughout the region. Figure 1 shows the most widely used terms and their correlation.

The sub-Ogallala surface.—Throughout the region, except in northern Nebraska, late Cenozoic deposits rest on an erosion surface developed on Mesozoic and Permian rocks, and therefore a record of Paleogene time is lacking. This Miocene erosion surface, offering but slight evidence of surface environment during earliest Neogene, was a late-mature topography of gentle slopes and broad valleys. Maximum relief probably did not exceed 250 feet, although it is judged that locally sharp incision occurred along minor tributaries. In the deeper parts of these Miocene valleys, the deposits that rest upon this erosion surface are commonly coarse clastics, and the soil profile developed on the bedrock has been removed; however, where the profile is partially preserved below the Ogallala it is relatively thin, is importantly influenced by the parent bedrock, and commonly lacks a zone of caliche accumulation.

The early Neogene drainage, generally aligned west to east, initiated sedimentation in the deepest parts of the major valleys and their principal tributaries (Frye, Leonard, and Swineford, 1956). Progressive reduction of topographic relief throughout the Neogene (fig. 2) by gradual engulfment

AGE			W. NEBRASKA	W. KANSAS	W. OKLA.	N.W. TEXAS
PLEISTOCENE	WISCONSINAN and RECENT	Almenan	(Alluvium and dune sand) Bignell	(Alluvium, loess, and dune sand) Bignell	(Alluvium and dune sand) Sanborn Formation	
		Bradyan	*Brady soil*	*Brady soil*		(Tahoka) (Peoria) (terraces)
		Scandian	Peorian Todd Valley	Peoria (sand and gr)		
	SANGAMONIAN		*Sangamon s.*	*Sangamon s.*		
	ILLINOIAN		Loveland Crete	Loveland Crete		"red cover sands"
	YARMOUTHIAN		*Yarmouth s.*	*Yarmouth s.*		
	KANSAN		Sappa Grand Island	Meade Sappa Grand Island	Meade Fm.	Tule Fm.
	AFTONIAN		*Afton soil*	*Afton soil*		
	NEBRASKAN		Fullerton Holdrege	Blanco Fullerton Holdrege	Blanco Fm.	Blanco Fm.
NEOGENE	(PLIOCENE)		OGALLALA GROUP Kimball Sidney Ash Hollow	Ogallala Formation Kimball Ash Hollow	Ogallala Formation	Ogallala Formation
	(MIOCENE)		Valentine	Valentine (Laverne)	Laverne Formation	

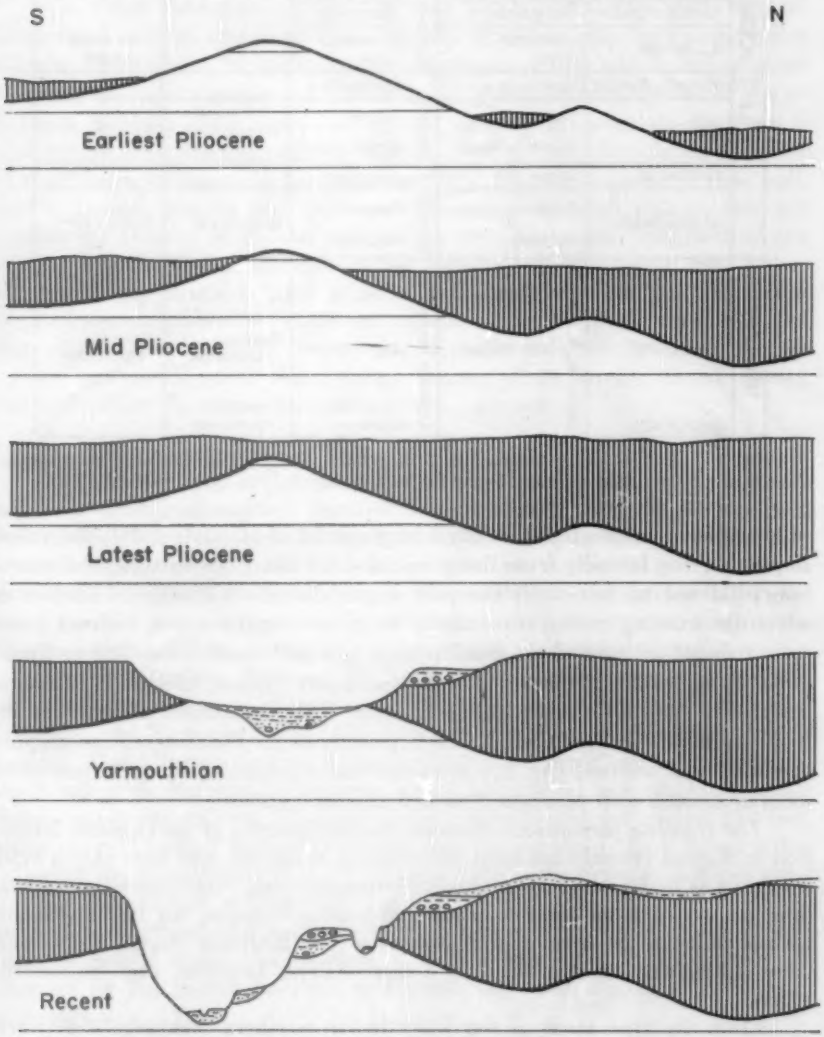
Fig. 1. Chart showing nomenclature in general use for late Cenozoic stratigraphic units in the central and southern Great Plains region.

of the erosional topography resulted from spread of alluviation onto the valley slopes. Moving laterally from linear initial areas along the streams, sedimentation produced an intricately complex degradational-constructional surface of which the existing record is primarily the depositional portion. Bedrock areas were reduced to isolated circumalluviated hills and small discontinuous linear areas along former major divides. This complex surface, graded to erosional surfaces in central Colorado and in the Flint Hills, maintained effective equilibrium in latest Neogene time. At the beginning of the Pleistocene, topographic evolution was reversed (fig. 2), and relief was accentuated by cyclic entrenchment of streams with resultant dissected alluvial terraces.

The Ogallala formation.—Because the stratigraphy of the Ogallala formation in Kansas recently has been described in detail, we give here only a brief summary as background for ecological interpretations. Stratigraphic classification has varied from place to place and among workers; for this discussion we use units in the sense of the Nebraska type localities (Lugn, 1939) and their equivalents as described in Kansas (Frye, Leonard, and Swineford, 1956).

From the type areas of the Valentine in northern Nebraska southward to central western Texas the initial deposits of the Ogallala consist generally of fine to medium sand and locally coarse gravel representing channel and near-channel floodplain deposits of major streams. Throughout the northern two-thirds of this region, the predominant color of these lowermost Ogallala de-

posits is greenish gray, and this fact, coupled with the general lack of carbonate cement, preservation of bedding, bentonitic alteration of volcanic ash lentils (Swineford, Frye, and Leonard, 1955), and preserved fauna and flora, indicates a continuously high water table in the floodplain areas. The Valentine member generally is well sorted fine to medium sand. An exception to this occurs in the panhandle area of Oklahoma and adjacent Kansas where



Schematic diagrams showing evolution of Central High Plains surface.

Fig. 2. Generalized diagrammatic cross sections showing evolution of the central and southern High Plains surface in Pliocene and Pleistocene time.

beds of diatomaceous marls and shales (Frye and Hibbard, 1941), locally classed as Laverne, reflect local structurally controlled ponding. This area also has afforded a record of mid-Neogene arboreal flora (Chaney and Elias, 1936) and exceptionally numerous molluscan fauna (Leonard and Franzen, 1944).

The Ash Hollow member conformably overlying the Valentine is comparable to it in thickness but much more widespread and continuous. By the end of deposition of the Ash Hollow, 80 to 90 percent of the erosional bedrock topography had been buried by these two members, which include most of the thickness of the Ogallala. Although gradational and generally separated arbitrarily by paleontological and paleobotanical evidence (Frye, Leonard, and Swineford, 1956), the two members can in general be distinguished by their gross lithology. The degree of sorting decreases in the transition zone between Valentine and Ash Hollow, and tones of red and red brown and zones of calcium-carbonate cementation appear. Upward into the Ash Hollow, degree of oxidation, irregularly spaced and discontinuous zones of calcium carbonate, strong alternation of channel and floodplain deposits, indications of developed soil profiles, and increased textural contrasts suggest a fluctuating, declining water table during deposition. As the surface was at all times an aggradational alluvial plain, a declining water table could not have been produced by physiographic changes but must have been related to climatic fluctuations.

The Kimball (including in its base the Sidney gravels of Nebraska classification) marks the top of the Ogallala formation and the culmination of the Neogene alluvial plain. Much thinner than either of the other members, it is more widespread, overlapping the Ash Hollow and at places resting directly on bedrock. The base is conformable and gradational except locally where it can be drawn sharply at the base of gravel lentils (Sidney). Depositional trends that were evident in the Ash Hollow are accentuated in the Kimball. Although the member is predominantly fine textured floodplain sediments, contained gravel lentils are exceptionally coarse; the colors, where not masked by caliche, are dark reds and red browns; bedding is almost nowhere preserved; indications of developed soil profiles are common; and the upper part of the member is everywhere characterized by conspicuous accumulations of calcium carbonate. These characteristics, accounted for only by a water-table position significantly below the surface, point to continued climate change in the late Neogene.

Pliocene-Pleistocene unconformity.—The relationships associated with the Pliocene-Pleistocene unconformity have obvious ecological implications. Neogene deposition culminated in a plain of low relief (fig. 2) marked only by natural levees, channel scars, and small low swells in local areas where the bedrock surface was not buried. This practically featureless surface maintained stability for a significant but unknown period of time, and indications of declining water table reached a maximum. A thick and widespread mantle of dense calcium carbonate, locally encrusted at the top, developed over most of the region. Although a variety of origins have been advanced, this mantle recently has been attributed directly or indirectly to soil-forming processes (Bretz and Horberg, 1949; Swineford, Frye, and Leonard, 1956). This hypothesis depends on a continued low water table with a balance between

precipitation and evaporation tending to reduce downward percolation of dissolved carbonates, followed by the removal (by wind, water, or both) of the friable A horizon. The solution effects and secondary encrustations now observable on this irregular and uneven limestone layer must have been produced by a subsequent episode of increased precipitation.

The equilibrium of this surface was interrupted with relative suddenness throughout the plains by what is judged to be the sharpest climatic change of the late Cenozoic. That stream incision was, at least locally, governed by climatic rather than tectonic factors is suggested by the fact that it took place along streams that rose on the plains surface as well as along those that flowed through the plains from the mountainous regions to the west. This initial episode of dissection, coming between the culmination of Ogallala deposition and the initiation of Blancan-Nebraskan deposition, is considered to mark the break between the Pliocene and Pleistocene.

The Pleistocene formations.—The depositional history of the Pleistocene contrasts strongly with that of the Neogene. In contrast to the progressive and gradually changing depositional trends of the Neogene, the Pleistocene is marked by comparatively shorter and more violent episodes of stream incision and alluviation (fig. 2) alternating with intervals of equilibrium recorded by widespread buried soils. Major stratigraphic units and their correlations (fig. 1) have been described in some detail (Condra and Reed, 1950; Evans and Meade, 1945; Frye and Leonard, 1951; 1952; Frye, Swineford and Leonard, 1948; Lugn, 1935; Schultz, Lueninghoener and Frankforter, 1951; Schultz and Stout, 1945; 1948).

The depth of initial Pleistocene incision varied widely, but, with the exception of some areas in Texas, the first episode of deposition (Blanco, Holdrege-Fullerton) was characterized by coarse gravels—generally coarser than the Ogallala deposits of the same region—grading upward into finer-textured clastics marked at the top by a lime-accumulating soil profile (Afton buried soil). In drainage ways originating in the plains, these sediments are largely reworked Ogallala and the coarser texture is due to removal of fines; but in through-flowing streams, Rocky Mountain-derived gravels locally are coarser than the coarsest elements of the adjacent Ogallala.

The episodes of stream incision following the Nebraskan were profound and general throughout the region. Kansan alluviation (Tule, Meade, Grand Island-Sappa) displays a coarse texture and is indicative of a degree of stream competence somewhat greater than that in the Nebraskan. A widespread fall of volcanic ash (Pearlette bed) occurred in the late phase of Kansan deposition and, although of slight ecological significance in itself, it trapped abundant molluscan faunas and furnishes a precise time datum throughout the plains (Frye, Swineford, and Leonard, 1948). Kansan sediments are widely preserved as terrace remnants. The Yarmouth soil that caps the sequence is deep and has strongly developed structure and caliche zone.

The Illinoian deposits (Crete-Loveland) at many places indicate a departure from the general cyclic pattern of the Pleistocene in the plains. Although there was local incision of the Kansan sediments, Illinoian deposits commonly overlie the Kansan with only minor incision, which suggests lack

of stream competence. This relationship also is true along drainage ways that originate within the plains, suggesting a decrease in the quantity of water available to the streams of the area. Moreover, the Illinoian is the first Pleistocene depositional cycle to have significant quantities of eolian deposits in upland areas. In the central and northern parts of the region the upland eolian deposits are largely "Loveland loess" and some local dune sand, whereas in the southern part extensive, well oxidized "cover sands" mantle many upland areas. The strongly developed Sangamon soil, readily traceable throughout the region, distinguishes the top of the Illinoian.

Renewed regional valley deepening preceded Wisconsinan deposition. This suggests increased stream competence, but as the Wisconsinan knickpoints have fallen far short of those reached by Kansan streams in the plains, the climatic implication is not clear. Furthermore, early Wisconsinan (Scandian) time witnessed the most extensive development of upland eolian deposition of the Pleistocene. Loess correlated as Peoria from northern Texas to northern Nebraska attained thicknesses of well over 100 feet, and extensive dune tracts formed in Texas, Kansas and Nebraska.

This major episode of Wisconsinan deposition was terminated by a brief interval of stability distinguished by the Brady buried soil. Post-Bradyan (Almenan) history is not consistent throughout the Plains. In the northern half, thin Bignell loess overlies the Brady soil at many places on the uplands, and sand dune tracts were developed; in valley areas post-Bradyan alluviation consists generally of a minor increment above the surface of the Brady soil followed by minor episodes of channel incision that have resulted in low narrow terraces and the present active floodplain. In the southern half of the region the Bradyan interval was marked by slight but general stream incision followed by minor alluviation that left the pre-Bradyan deposits standing as a low terrace. Here in upland areas, eolian activity was limited to local shifting of fine-textured materials, mostly from nearby blowout areas.

Early and late Wisconsinan water-laid deposits present a textural gradation, as do the older cycles, but on the average they are finer. Subsequent to the Bradyan, several minor soils formed and were locally buried, indicating a complex of minor depositional cycles.

ECOLOGICAL TRENDS

The succession of gross ecologies of the Great Plains through late Cenozoic time becomes reasonably clear when we add to the history of sedimentation the data from the almost continuous sequence of populous molluscan faunas, the distinctive floras of the Neogene, and climatic and floral implications of the buried soils. These data are summarized and environmental interpretations are shown diagrammatically in figure 3.

The rigors of the present climate have largely exterminated the rich and varied faunas of gastropods that thrived on the Great Plains during one part or another of late Cenozoic time, but most of the species, or their near relatives, still live somewhere in North America. This fact makes possible accurate interpretations of past ecologies in terms of present environmental requirements of these animals. In general the same may be said for Tertiary fossil

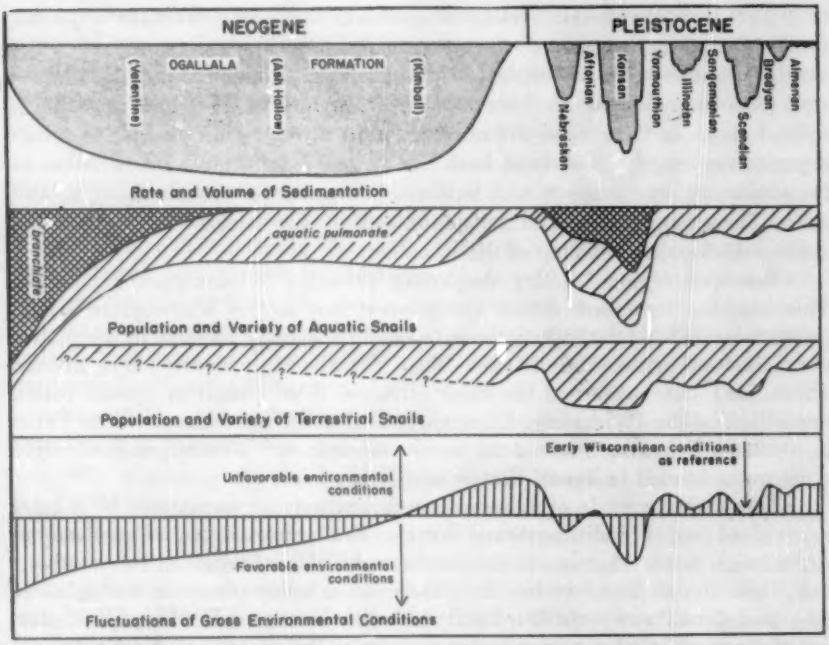


Fig. 3. Schematic diagrams showing ecological trends in the central Great Plains during late Cenozoic time. The curves depicting sedimentation, and population and variety of moluscan faunas, are generalities for the region (Texas to Nebraska) and are not intended as an accurate picture for any specific locality.

plants (Elias, 1942; Frye, Leonard, and Swineford, 1956), but plant remains in Pleistocene deposits in this region are too rare to have any practical utility.

At the initiation of Ogallala deposition, in approximately mid-Neogene time, organic remains confirm the existence of somewhat moist conditions. Western Oklahoma has furnished particularly abundant materials from the Laverne beds (lower Valentine member). Plentiful and stable water conditions are required by such branchiate gastropods as *Amnicola*, *Calipyrgula*, and *Viviparus*, which lived in association with species of the aquatic pulmonates *Lymnaea*, *Helisoma*, *Gyraulus*, and others. The terrestial gastropods that occur in this assemblage (*Pupoides albilabria*, *Vertigo ovata*, *Hawaiia minuscula*, and others) are largely wide-ranging species that have persisted on the plains to the present in spite of climatic fluctuations. The presence in these earliest Ogallala deposits of leaves of the persimmon (*Diospyros pretexana*) (Chaney and Elias, 1936) seems to point to a somewhat higher average annual temperature, but this may be only the effect of a more stable climate.

Ascending the stratigraphic column to the transition zone at the top of the Valentine member and into the Ash Hollow member, the evidence of declining water-table and progressively irregular stream flow is augmented by indications of declining and less stable surface moisture conditions. Branchiate gastropods, prevalent in the Valentine, disappear in the Ash Hollow. In the

upper Valentine member the flora is dominated by species of prairie grasses (*Stipidium commune, S. coloradense, S. breve, S. elongatum, S. kansasense, Berrichloa amphoralis, B. conica, B. tuberculata,* and others) and forbs (*Krynitzkia coroniformis, K. auriculata, Biorbia fossilia*), which replace the arboreal vegetation that was at least locally common in the early Neogene. These floral elements change through the remainder of the Neogene, but *Biorbia fossilia* and the single tree *Celtis willistoni* persist well into the Kimball, where they are associated only with *Prolithospermum johnstoni* and a single species of *Stipidium*. In the upper Ogallala deposits the gastropod fauna is likewise sparse and is composed almost entirely of species of *Physa, Helisoma,* and *Gyraulus,* which are capable of surviving long periods of desiccation.

The organic remains in Neogene strata corroborate the interpretations reached from the clastic sediments. Progressive desiccation of surface moisture conditions and deterioration of the mild and tolerant climate of the early Neogene is clearly denoted by disappearance of the local arboreal flora, followed by deterioration of the succeeding prevalent lush prairie flora, and accompanied by gradual disappearance of the large and diversified assemblage of branchiate gastropods and eventual elimination of many of the aquatic pulmonate forms.

Organic remains are not available to furnish climatic evidence for the period of virtual equilibrium that forms the erosional-depositional hiatus between the Pliocene and Pleistocene. However, because the climatic trends of the Neogene seem orthogenic and the lithologic and geomorphic evidence points to this interval as the culmination of the trend toward aridity, it is judged that conditions on this stable surface were at least as dry and unfavorable to organic life as are those of the present. In fact, it seems likely that the generally featureless surface, low regional water table and resultant scarcity or lack of permanent streams, and the almost universal surficial blanket of caliche coming at the end of a long period of increasing dryness denote the most adverse conditions to plant and animal life of any time during the Cenozoic.

A distinct climatic reversal characterizes the beginning of the Pleistocene. This is indicated not only by the geomorphological and stratigraphic evidence but also by the reappearance of branchiate snails and a marked increase in the abundance and kinds of the total gastropod fauna in the plains.

The reversed climatic trend toward moist and tolerant conditions reached its climax during Kansan time. The deep stream entrenchment and coarse-textured deposits throughout the plains was accompanied by a resurgence of a varied branchiate gastropod assemblage (Leonard, 1950). *Amnicola,* which first reappeared in Nebraskan deposits, was joined by *Valvata* as a dominant form, along with a wide variety of aquatic pulmonates including *Physa, Helisoma, Gyraulus, Menetus, Ferrissia,* and *Lymnaea.* Terrestrial gastropods point to a dominant prairie vegetation, associated with belts of trees and shrubs along valleys. The fauna included *Hendersonia occulta,* several species of *Gastrocopta, Vertigo,* and *Pupilla, Retinella indentata* and *R. electrina, Stenotrema leai, Strobiliops sparsicosta, Euconulus fulvus,* several species of *Succinea,* and other terrestrial gastropods of less importance. These terrestrial

gastropods are species characteristic of the prairies or forest borders, and it is noteworthy that the large "helicine" gastropods, such as *Polygyra, Anguispira, Triodopsis, Mesodon,* and *Allogona,* which flourish in the forests a few hundred miles east of the Great Plains, do not occur. The buried Yarmouth soil profile (Frye and Leonard, 1952, 1954) reflects development under a grasslands cover.

Following the Kansan, a strong, distinctive, but oscillating trend toward increasing aridity in the Great Plains continued into Recent time. During the Yarmouthian, branchiate gastropods disappeared from the plains, and terrestrial gastropods declined in number, kinds, and population density (Leonard, 1952). These faunal changes support the strong physical evidence of pronounced climatic deterioration by Illinoian time.

The extensive upland deposits of early Wisconsinan sediments have preserved the buried Sangamon soil that has been studied at several hundred localities throughout much of the region. This buried soil furnishes a type of regional data not available at other stratigraphic positions in the late Cenozoic. When it is compared with the present surface soils in the central Great Plains, from positions of similar drainage conditions and parent material it becomes clear that the belts of the great soil groups are offset to the west by as much as 100 miles, which verifies the conclusion that the Sangamonian interval was more moist than the present. The thick and massive zones of caliche accumulation in the southern part of the region, however, make a regional westward offset in soil groups less clear, raising the possibility that here climatic conditions in the Sangamonian approached those of the present.

In pre-Bradyan Wisconsinan (Scandian) time the trend toward an arid climate halted temporarily. The aquatic pulmonate snails *Physa, Helisoma,* and *Gyraulus* were widespread, and the most common terrestrial snails were species of *Vallonia, Pupilla, Gastrocopta, Pupoides,* and *Succinea.*

The climax of the climatic trend toward the present semiarid condition of the plains was reached during the Bradyan interval or in the early part of succeeding Almenan time. The Brady buried soil shows no westward offset from the present surface soil, and Almenan molluscan faunas are comparable to those living in the region today. In Almenan deposits, only *Hawaiia minuscula, Succinea avara,* and *Pupoides albilabris* are widely distributed. These hardy species, capable of enduring extremes of summer heat and winter cold, together with the aquatic pulmonate snails *Helisoma trivolvis, Physa anatina,* and a few *Gyraulus,* are all that remain to form the general molluscan fauna of the Great Plains. Other species do occur but always in extremely local and abnormal ecological situations. It is obvious that the present rigorous climate of the Great Plains was initiated during the Bradyan interval.

Although the climate of the Recent is judged to approach that of latest Tertiary time, there are certain significant differences between the gross ecological situations. Now there exists a distinct and sharp topographic relief along major valleys, and although the water table under the uplands may be much lower than in latest Tertiary times, it is nevertheless sufficiently high to maintain as perennial the streams in these deeply incised through valleys. Furthermore, existing physiographic conditions produce relatively strong local

variations in ecological conditions that could not have existed on the almost featureless plain of alluviation in latest Tertiary time.

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SEDIMENTARY CHARACTERISTICS OF DUST STORMS: I. SORTING OF WIND-ERODED SOIL MATERIAL

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ABSTRACT. Sorting of soil materials by the wind is an intricate phenomenon. The most distinct feature in the whole sorting process was found to be the peak diameter of the saltating grains. Fractions larger than the peak diameter tend to remain in the wind-eroded fields, and particles smaller than this diameter tend to be deflated and carried far through the atmosphere. Depending on soil class, from 31 to 78 percent of particles smaller than 0.1 mm in diameter contained in the wind-transported soil fraction are deflated by a single windstorm. Silt generally is more readily deflated than sand or clay. Wind erosion has caused little change in texture of loess soils but has tended to remove the fine constituents from the coarser-textured soils, leaving the sand behind. This sorting process if continued even for a day or two adds considerably to the general sandiness of the affected areas and to consequent irreparable depletion of soil productivity.

INTRODUCTION

An eroding wind acts on the soil like a fanning mill on grain—removing the finer and more porous particles and leaving the coarser and denser behind. There is a wide range of rates of removal of different soil fractions. Some fractions are not moved at all; other fractions are moved slowly by surface creep; some are transported rapidly in saltation; and still others are carried through the air in suspension. A substantial amount of study has been made by Bagnold (1943) on the grading of sand by wind, but only preliminary information is available on the sorting action of wind on arable soils (Fly, 1935; Daniel, 1936; Chepil, 1946). A series of studies was undertaken, therefore, on the sorting of various soils by wind, on the atmospheric concentrations of dust and rate of dust removal from present-day wind-eroded regions, and on the composition of dust originating from different soils. The first of these was on the intricate phenomenon of the sorting of soil materials by the wind and on effects of the sorting process with respect to possible changes in the productivity of agricultural lands. The results of this study are reported herein.

PROCEDURE

Samples of soil when in a dry condition were taken to about one-inch depth from wind-eroded fields, from coarse eroded materials deposited in surface depressions of the fields, and from newly accumulated drifts within or near the fields. Five individual samples collected from different parts of each field were combined, oven-dried at 175°F, and thoroughly mixed. The samples of drifted material represented that portion of the soil moved primarily in surface creep and saltation and deposited in or near various obstructions or traps such as vegetation and ground depressions during a single windstorm. The individual samples were taken from different positions on the drifts to assure a composite representing the average. Sampling was conducted during the spring seasons of 1949, 1950, 1954, and 1955 when erosion of soil by wind was prevalent. The samples were collected from central and western Kansas and eastern Colorado. A total of 44 wind-eroded fields having a soil class ranging from sand to clay were included in the study.

Determinations of the physical and chemical soil characteristics included (a) the size distribution of dry particles and aggregates by the method of Chepil (1952), (b) the bulk density of dry particles by the elutriation method of Chepil (1950), (c) the size distribution of water-stable particles and aggregates by the modified method of Yoder (1936), (d) the mechanical composition by the method of Bouyoucos (1951), and (e) the organic matter by the method of Walkley (1947).

For the sake of simplicity only average values for each soil class are included in this paper.

THE NATURE OF ELUTRIATION

The size distribution of dry particles and aggregates in soil material of wind-eroded fields and of soil drifts originating from and trapped in the vicinity of the fields is shown in figure 1. Because the proportions of the finest grades in soil materials from wind-eroded fields and from the drifts was low but important compared with the proportion of the predominant grades, it was decided to follow the method of Bagnold (1943) for a graphic representation of the size distribution of these materials. This is done by plotting the logarithm of the percentage weight of soil, Δp, per unit of log-diameter scale against the logarithm of grain diameter. Assuming the diameter limits of each grade as determined by sieving to be d_1 and d_2, the logarithmic interval of grain diameter is given by $\log d_1/d_2$ which, for convenience, is designated by ΔD. The percentage weight of each grade per unit of log-diameter scale is then equal to $\Delta p/\Delta D$.

The soil material of wind-eroded fields (shown by hatched lines in figure 1) had either two or three diameter peaks, depending on soil class. The sand, loamy sand, and silty clay and clay had two peaks, one centered around 0.3 mm for sandy soils and around 0.6 mm for clayey soils and the other at about 25 mm for sandy soils and 10 mm for clayey soils. The intermediate-textured soils, on the other hand, had three distinct diameter peaks, one occurring at about 0.07 mm, another between 0.3 and 0.6 mm, and the third between 10 and 30 mm.

In soil material taken from the drifts (shown by continuous lines in figure 1) there was only one peak diameter. The peak was at either 0.4 or 0.6 mm depending on soil class. The peak diameter of the drifted material derived from fields composed of sand and loamy sand was about 0.4 mm and of drifted material from the finer-textured soils the peak was about 0.6 mm. The reason for the difference in the peak diameter was quite apparent. The drifted materials derived from fields of sand and loamy sand were composed principally of discrete, non-porous grains having an average bulk density of 2.37. The materials drifted from the finer-textured soils, on the other hand, were predominantly aggregates exhibiting a distinct degree of porosity and having an average bulk density of 1.70 (table 1). Their equivalent peak diameter, therefore, was equal to $\sigma d/2.65$ in which σ is the bulk density of a discrete soil grain or aggregate of diameter d. In accordance with the above expression, the average equivalent peak diameter of grains eroded from these soil classes was also about 0.4 mm. The occurrence of a similar equivalent

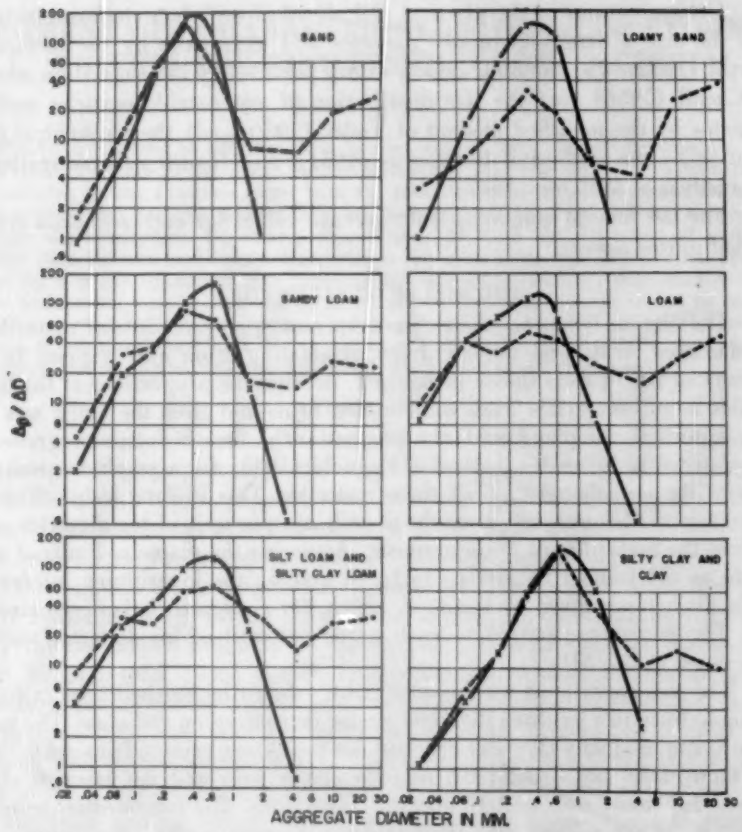

Fig. 1. Size distribution of dry aggregates in surface inch of soil material from wind-eroded fields (shown by hatched lines) and from soil drifts originating from and trapped in the vicinity of the fields (shown by continuous lines) for 6 textural groups of soils. The number of fields represented in the 6 textural groups, starting with the coarsest group, was 2, 3, 8, 12, 14, and 3.

peak diameter of erodible particles in all soil classes herein investigated is to be expected since the natural wind forces operating on all soils of the region were about the same.

The diameter peak centering about 0.07 mm in soil material of wind-eroded fields apparently caused a slight bulge in drifted material of the same diameter. The second peak in the soil material of wind-eroded fields coincided approximately with the peak diameter of the drifted materials. In no case, however, was the third peak in soil materials of the wind-eroded fields reflected in the composition of the drifted materials. The reason for this is clear. The size indicated by the third peak was definitely non-erodible and therefore did not influence the composition of the drifted materials. It is evident from figure 1 that some sizes on the extreme right side of the peak diameter of drifted materials were not moved by wind. The next smaller sizes exhibited some

TABLE 1
Average Bulk Density of Discrete Particles in Various Dry Sieve Fractions of Various Soils

Soil	Particle bulk density in sieve grade					
	2.0-.84 mm	.84-.42 mm	.42-.21 mm	.21-.1 mm	.1-.05 mm	<0.05 mm
Sand	2.59	2.56	2.57	2.53	2.51	2.49
Loamy sand	2.09	2.11	2.31	2.40	2.16	2.11
Sandy loam	1.57	1.61	1.59	1.63	1.79	1.97
Loam	1.52	1.56	1.69	1.69	1.72	2.02
Silt loam and silty clay loam	1.57	1.57	1.63	1.64	1.78	1.82
Silty clay and clay	1.61	1.66	1.71	1.70	1.79	2.02
Grand average	1.82	1.84	1.92	1.93	1.96	2.07

movement, but their movement was apparently slow and low in magnitude. These fractions constituted the bulk of the lag materials moved primarily by surface creep. The next smaller fractions were centered around the peak diameter of materials found in drifts or dunes. They constituted grains moving primarily in saltation. Particles still smaller than these, shown on the left of the peak diameter, contained what may be termed the "dust" fraction. The dust particles moved rapidly, and a large proportion of them, especially on the extreme left of the peak diameter, were lifted in the air and carried away in what may be considered true suspension. However, there was no clear-cut demarcation where one soil grade ended and the other began. Transition from one grade to another was gradual and, in fact, one grade merged considerably with the other. One reason for this sort of overlapping was that some particles in surface creep moved intermittently in saltation and some of those in saltation no doubt were lifted by gusts of wind and carried in suspension. Even for a given type of movement the rate of movement of the individual soil particles was highly variable due to variation in size, shape, and density of the particles and to variation in the force of the wind.

An example of size distribution of particles in drifts and lag materials formed by unusually strong winds is given in figure 2. The peak diameter of particles in the lag materials was about twice as large as in the drifts. Same as particles in drifts, the lag materials were characterized by a single peak diameter with arms to the left and the right of the peak falling off independently at a more or less uniform rate.

MAGNITUDE OF SORTING OF ERODED MATERIALS

Let it be assumed that a and b are the percentages of the sieve fraction in drifted material and in field soil, respectively, which have a grain diameter corresponding approximately to the peak diameter of the drifted material (fig. 1 and table 2). If the proportion of the next smaller sieve fraction c to fraction a in the drifted material is $\frac{c}{a}$ and the proportion of the same respective sieve fractions in the field soil is $\frac{d}{b}$, the percentage loss of fraction

TABLE 2

Average Size Distribution of Dry Particles and Aggregates of Surface Inch of Soil Material from Wind-eroded Fields and from Soil Drifts Originating from and Trapped in the Vicinity of the Fields

Source	Dry sieve fractions								
	>12.7 mm	12.7-6.4 mm	6.4-2.0 mm	2.0-.84 mm	.84-.42 mm	.42-.21 mm	.21-.1 mm	.1-0.05 mm	<0.05 mm
	%	%	%	%	%	%	%	%	%
Fields	8.7	6.7	6.1	8.4	21.9	25.2	11.0	7.2	4.8
Drifts	0	0.1	0.4	7.2	35.8	34.2	14.0	5.7	2.6

c from the drifted material is $100\left(\frac{d}{b}-\frac{c}{a}\right)/\frac{d}{b}$. Similarly, losses from the drifted soil of other sieve fractions on each side of the peak diameter can be determined. Losses of sieve fractions smaller than the peak diameter indicate the approximate proportions of various sizes of dust particles lost to the atmosphere, and losses of sieve fractions larger than the peak diameter indicate the approximate proportions of lag and non-erodible fractions which remained

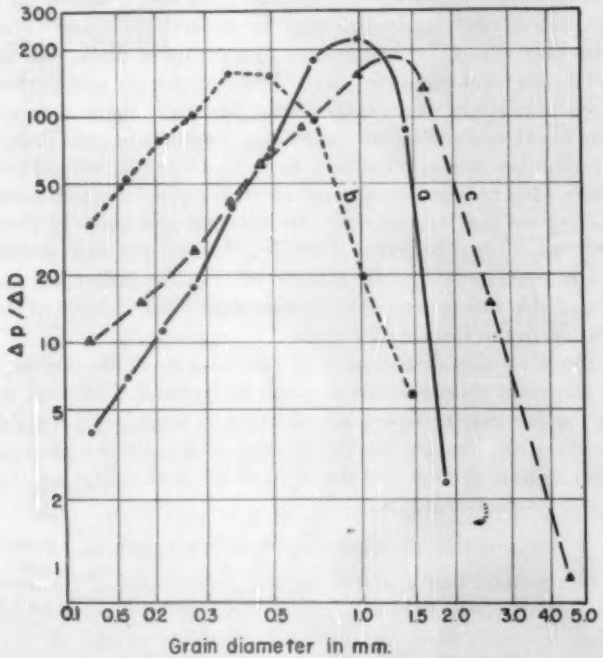

Fig. 2. Size distribution of soil grains moved by unusually high winds: (a) Lag material of average grain bulk density of 1.80 deposited in a 2-foot depression by a wind of 50 mph at a 5-foot height on March 24, 1949, near Salina, Kansas. (b) Material jumped across the depression and piled in drifts against a highway. (c) Lag material of average grain bulk density of 1.83 deposited in small gravel ridges by a wind of about 70 mph at a 5-foot height on February 19, 1954, near Tribune, Kansas.

in the field and from among which the highly mobile fractions were removed. Since the drifted materials accumulated usually in small dunes or drifts throughout the eroded area, they contributed appreciably to the general soil composition of the area. The dust raised in suspension, on the other hand, was carried far and wide and constituted that portion of the soil material lost from the wind-eroded region. The removal of dust by suspension is known as deflation.

In accordance with the above method of estimation, the approximate proportions by weight of dry dust particles lost from the various erodible sieve grades by the process of deflation are shown in table 3. These data show that the smallest soil particles deflated the most, whereas the particles approaching the peak diameter of the drifted material deflated least. Depending on soil class, from 32 to 84 percent of particles smaller than 0.05 mm in diameter and from 30 to 72 percent of particles between 0.05 and 0.1 mm contained in the wind-transported soil fraction were deflated by a single windstorm. The degree of deflation, i.e., the percentage of total quantity of dust deflated by wind, was least from silty clay and clay soils (table 3). More-

TABLE 3
Average Proportion of Wind-eroded Soil Lost from Various Sieve Grades by the Process of Deflation

Soil class of eroded fields	Proportion removed from sieve grades				
	0.84-0.42 mm	0.42-0.21 mm	0.21-0.1 mm	0.1-0.05 mm	<0.05 mm
	%	%	%	%	%
Sand	0	0	23	50	50
Loamy sand	0	0	12	55	78
Sandy loam	0	47	56	72	84
Loam	0	0	3	53	69
Silt loam and silty clay loam	0	6	14	56	74
Silty clay and clay	0	1	7	30	32
Grand average	0	9	19	53	64

over, these soils contained the lowest proportion of the dust fraction (table 2). Consequently the magnitude of deflation (based on degree of deflation multiplied by the proportion of dust contained in the wind-eroded material) was least on silty clay and clay. On the other hand, the intermediate-textured soils, i.e., loam, silt loam, and silty clay loam indicated a high degree of deflation, contained the greatest proportion of dust subject to deflation, and therefore had the greatest magnitude of deflation.

EFFECTS OF SORTING ON SOIL COMPOSITION

In virtually all soil classes investigated, the drifted materials contained much less fine and more coarse water-stable particles than the field soil (fig. 3). Apparently a considerable proportion of the fine water-stable particles in that portion of the soil moved by wind was removed by deflation. The coarse

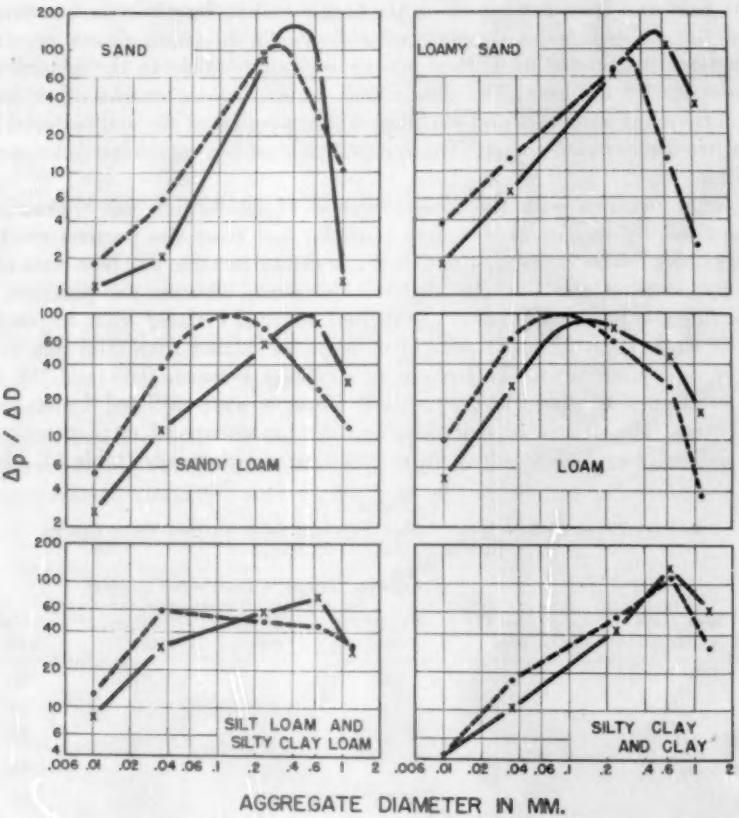

Fig. 3. Size distribution of water-stable aggregates in surface inch of soil material from wind-eroded fields (shown by hatched lines) and from soil drifts originating from and trapped in the vicinity of the fields (shown by continuous lines) for various textural groups of soil.

water-stable particles or aggregates, on the other hand, tended to accumulate in the drifts. Unlike the secondary dry aggregates or clods, the water-stable particles or aggregates appeared to be extremely resistant to disintegrating forces of wind erosion. Unfortunately, virtually all of them in each of the soil classes investigated were too small to resist movement by wind. There were apparently two distinct phases of the sorting of water-stable particles by the wind: (a) the removal from the drifted portion of one-third to two-thirds of the particles smaller than about 0.1 mm in diameter, and (b) accumulation into drifts of almost all of the remaining water-stable fractions.

The proportion of elementary mechanical fractions, sand, silt, and clay, was influenced considerably by the sorting action of wind erosion (table 4). The nature of the sorting action was not uniform in all soils; it varied appreciably with soil class. But there were some aspects of the sorting action that were present in all soils. In virtually all cases, for example, the proportion

of sand increased, and in all cases the proportion of silt decreased in that portion of the soil that was moved and deposited in drifts by the wind. In the drifted materials, the original content of silt was reduced from 7 to 67 percent, depending on soil class. These reductions were statistically significant in all soil classes except sand, loamy sand, and clay. Lack of significant differences in at least some of these cases was apparently due to the fact that not more than two or three fields were chosen in each case.

The relative proportions of clay in the drifts and in the wind-eroded fields varied depending on soil class. All sandy soils had substantially lower proportions of clay in the drifts than in the fields, but all of the finer-textured soils had a somewhat higher proportion of clay in the drifted materials than in the field soils. In these latter soils the clay fraction apparently tended to stick to those particles or aggregates that were large enough to remain in the drifts in the vicinity of the eroded area.

The remarkable changes in mechanical composition of the soil fraction that was moved by wind were a result of a single windstorm! In virtually all sandy soils a storm was sufficient to change the texture of the wind-transported soil material from a loam to a sandy loam, or a sandy loam to a loamy sand, or a loamy sand to a sand. In finer-textured soils, on the other hand, such as in loessal soils, the sorting was far less pronounced. From these soils literally all of the three mechanical constituents moved by wind were lost substantially to the atmosphere, although the clay fraction appeared to be somewhat less mobile than the other two. Mobility here is interpreted not as the speed of

TABLE 4

Mechanical Composition and Organic Matter of Surface Inch of Soil Material from Wind-eroded Fields and from Soil Drifts Originating from and Trapped in the Vicinity of the Fields

Soil class	Source	Mechanical composition			Organic matter
		Sand >0.05 mm	Silt 0.05-0.002 mm	Clay <0.002 mm	
		%	%	%	%
Sand	Fields	91.8	3.3	4.9	0.44
	Drifts	97.2	1.8	1.0	0.30
Loamy sand	Fields	80.5	11.3	8.2	0.63
	Drifts	90.8*	6.2	3.0	0.44
Sandy loam	Fields	64.8	23.0	12.2	1.19
	Drifts	84.9**	7.7**	7.4*	0.62*
Loam	Fields	36.8	45.6	17.6	1.78
	Drifts	50.2**	31.9**	17.9	1.62
Silt loam	Fields	18.4	58.2	23.4	2.41
	Drifts	17.0	53.4*	29.6**	2.53
Silty clay loam	Fields	22.2	49.1	28.7	2.27
	Drifts	24.4	45.6*	30.0	2.35
Silty clay and clay	Fields	11.8	35.1	53.1	2.41
	Drifts	13.3	31.3	55.4	2.76

* Difference significant at 5 percent level by "F" test.
** Difference significant at 1 percent level by "F" test.

the particles through the air, but as their relative rate of removal from the soil.

One other soil element affected by the sorting action of the wind was the organic matter (table 4). Again, the effects of the sorting action varied, depending on soil class. In all sandy soils and in loam the organic matter content of the accumulated soil drifts was substantially lower than in the residual soil of the windblown fields. Much of the organic matter apparently was not bonded with the coarse soil fractions and was deflated readily by the wind. In loessal and fine alluvial soils the organic matter content of the accumulated soil drifts was somewhat higher than in the field soils. The organic matter in these soils apparently tended to be associated with that mineral portion of the soil that was not deflated by the wind. In addition, there was some accumulation of finely divided vegetative residue in the drifts. This was not so in sandy soils. Here the drifts as well as the residual soil of the windblown fields contained much less organic matter than the soil in non-eroded fields. Much of the organic matter in these soils apparently was carried into the atmosphere and was thus completely removed from the wind-eroded areas.

CONCLUSIONS

The wind erosion process was composed of two major phases: the sorting of various primary and secondary soil fractions and the disintegration of the secondary fractions to the primary particles by the abrasive action of wind-eroded grains.

In the first phase, the wind-eroded soil material tended to be sorted out into several distinct grades some of which were as follows: (1) the non-erodible fractions remaining in their original locations in the field, (2) the so-called "lag" materials moving slowly by surface creep and deposited primarily in surface depressions throughout the field, (3) the grains moving rapidly in saltation and deposited usually in drifts or mounds throughout the affected area, and (4) the dust particles kicked up by the saltating grains to form dust clouds usually carried great distances from their source. The data indicated that removal of dust particles by deflation, although variable in magnitude, was appreciable for each soil investigated. On the average, almost 60 percent of the total dust content (particles smaller than 0.1 mm) in that portion of the soil moved by a single windstorm was deflated.

In the second phase of the wind erosion process, both the transported and the stationary soil fractions were disintegrated to smaller fractions by the abrasive action of the transported grains. The impacts of grains in saltation appeared to cause the greatest degree of abrasion. The secondary aggregates or clods tended to break down usually into primary (water-stable) aggregates most of which were readily transported by the wind. These primary aggregates in turn tended to break down into individual mechanical fractions—sand, silt, and clay.

Silt was more readily deflated than either sand or clay, except from soil classes such as sand and loamy sand already having a relatively low ratio of silt to clay. The silt particles exhibited little cohesion and were therefore readily separated by disintegrating forces of wind erosion (Chepil, 1955).

Moreover they were small enough to be readily deflated by the wind. The clay particles, on the other hand, exhibited great cohesive property and showed a definite tendency to form stable aggregates too large to be lifted into the atmosphere.

The sorting process of wind erosion has caused a profound change in the mechanical composition of some soils and little change in others, depending on their physical characteristics. On fine alluvial and loessal soils composed mainly of silt and some fine sand and clay, the sorting process of wind erosion did not appreciably change the elementary mechanical composition of the residual soils. Here the wind had no great tendency to sort out the various mechanical constituents. The sorting action was limited particularly to separation of the different sizes of the secondary particles or aggregates which apparently contained a more or less uniform proportion of each of the mechanical constituents.

In contrast, a considerable degree of sorting of elementary mechanical fractions occurred in loam and sandy soils. These soil classes contained a considerable proportion of sand too coarse to be deflated by the wind. Consequently, this sand fraction tended to remain in the drifts while much of the silt and clay was lost to the atmosphere. The drifts were deposited here and there within the affected area and added that much more to the general sandiness of the area. In addition to the reduction of organic matter and consequently of soil productivity, the depletion of silt and clay and the accumulation of sand added further to the hazard of wind erosion and to the problem of how to hold the remaining soil. The sorting of mechanical fractions such as occurred on these soils constitutes one of the most serious aspects of the wind erosion problem.

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SILICIFIED SEDIMENTARY ROCKS IN AUSTRALIA

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ABSTRACT. Three types of surface rock are described: (a) silicified material resembling certain South African silcretes in occurrence and petrography; (b) silicified aeolian sands; (c) sub-basaltic quartzite exposed by removal of its igneous cover. The rocks contained authigenic quartz, chalcedony, and opal, all of which corroded clastic quartz. The difficulties in the nomenclature and diagnosis of quartz, chalcedony, and opal are discussed. Siliceous veins in (a) originated by dilation or replacement. Replacement was of one texture by another rather than of one chemical compound by another, i.e. the veins were the paths along which the passage of solutions had caused recrystallization of silica. There were analogies with phenomena shown by permeation gneisses. The origin of the solutions responsible for the silicification of (a), (b), and (c) is discussed. Attention is directed to the possibility that former silcrete horizons are preserved among the rocks of the stratigraphical record.

INTRODUCTION

The silicified rock common at the surface in the more arid regions of Australia is a component of the "duricrust" (Woolnough, 1927). It has been dubbed "grey billy," but Waterhouse and Browne (1929) remarked that rock silicified beneath basalt flows had received the same name. Some silicified surface rock is termed "porcellanite" in Australia, i.e. when its texture has some resemblance to that of unglazed ceramic artefacts. In American publications "porcellanite" has been applied to bedded sedimentary rocks, to cherts, and even to lithified volcanic glass dust (Williams, Turner, and Gilbert, 1954, p. 154, 359, fig. 124c). Tomkeieff (1940), writing in a British journal, used it to designate non-vitreous contact-metamorphosed rocks more minutely crystallized than hornfels; he listed yet other meanings that had been attached to it in the past. The name "silcrete," applied by Lamplugh (1907) to silicified surface rocks in southern Africa, although not current in Australia, is appropriate for some Australian materials.

The Australian rocks that may be appropriately called "silcrete" resemble their South African analogues in mode of occurrence, chemical composition, and petrographic detail. Both have tended to develop in peneplaned regions of sporadic rainfall, i.e. where the effects of chemical weathering were prolonged and cumulative but the removal of the products of rock decay was tardy. In some areas subsequent denudation undermined the silicified crust and scattered its fragments to contribute to the Australian "gibber plains" (David and Browne, 1950, pl. 54c) or the "silcrete rubble" of South Africa. Where silcrete rubble was resilicified it became "silcrete conglomerate" (Frankel, 1952). Such conglomerate is evidence of more than one period of silicification; certain Australian surface rocks provide similar evidence.

Silcretes have been called "surface quartzites" (Frankel and Kent, 1937), but it is now known that they can occur not only at but beneath the surface. Thus they may be intercalated in the clay formed by the weathering of shale or tillite (Mountain, 1952; Frankel, 1952). Leached clays are frequent beneath the Australian duricrust (Woolnough, 1928), but it is not yet known if they contain silcretes.

Published studies indicate that the common types of South African silcrete originated (a) on or in clay formed by the weathering of shale or

tillite; (b) in soils or subsoils derived from sandstones or pre-existent silcretes; (c) at the exposed surfaces of undisintegrated sandstones. All three types of silcrete were formed from material silicified in place but they may associate with silicified transported material, e.g. gravel, screes, or aeolian sand; alternatively, the latter material may occur independently (Frankel and Kent, 1937; Frankel, 1952). Silicified transported material, e.g. aeolian sand, is to be seen also in Australia, but this Continent provides in addition numerous instances of rocks silicified beneath the widespread Tertiary basalt flows. Among the affected rocks were sand mantles derived from underlying sandstones (Waterhouse and Browne, 1929) and river sands and gravels (Raggatt, 1938).

Silcretes cannot be expected to be peculiar to South Africa and Australia; they should occur wherever topographic and meteorologic conditions have been appropriate. Moreover, their parental rocks need not be sedimentary. Thus serpentines were among the silicified surface rocks noticed in Cuba (Flint, 1952), and some South African silcretes were derived from dolerites (Frankel and Kent, 1937; Mountain, 1952).

The possibility that "fossil" silcrete horizons may occur among the rocks of the stratigraphical record is one to which attention may be directed.

THE SCOPE OF THE PRESENT STUDY

Most of the specimens were collected during a reconnaissance designed to find new or better sources of raw material for the manufacture of silica bricks. Silcretes were especially sought because there is a considerable production of silica bricks from them in South Africa, and these bricks have properties so desirable that South African silcrete has been imported by British manufacturers (Davies, 1952).

The rocks to be described are classified for convenience as silcretes, cemented aeolian sands, and sub-basaltic quartzites; the cemented sands may, alternatively, be regarded as a variety of silcrete (Lamplugh, 1907).

All three rock types contained one or more of the minerals commonly described as quartz, chalcedony, or opal, but the diagnosis was not always free from ambiguity, especially as there is some doubt as to how far these minerals represent distinct forms of silica.

THE RECOGNITION OF QUARTZ, CHALCEDONY AND OPAL

Quartz of adequate crystal size, and fibrous chalcedony, are both readily identifiable under the microscope, but cryptocrystalline silica, not manifestly fibrous, may prove difficult to refer to either of these minerals. A refractive index below that of quartz suggests chalcedony but may be caused merely by micropores (Midgley, 1951; Weymouth and Williamson, 1951). Such material may show an X-ray diffraction pattern not obviously different from that of quartz and this, indeed, is true also of typical fibrous chalcedony.

Microscopical identification of opal is based usually on the isotropic character and low refractive index; birefringence does not invalidate the diagnosis, as this can be variously attributed to the effects of strain in an amorphous material or to the development of a degree of crystallinity. Levin and Ott (1933) showed that opal may contain quartz or high or low cristoba-

lite, and this, according to Dwyer and Mellor (1933, 1934) was true also of opals from Oodnadatta and Coober Pedy, i.e. localities near which were collected the silcretes described below.

Buerger (1951), however, found that opal shows a continuous transition from apparently amorphous material to well crystallized high cristobalite as the size of the constituent units increases. By appropriate heat treatment Endell (1948) obtained silica intermediate in crystallinity between "amorphous" kieselguhr and low cristobalite. These findings have some analogy with those of Buerger.

Isotropism under the polarizing microscope can be exhibited not only by opal but by silica with both the refractive index and X-ray diffraction pattern of quartz (Winchell, 1924). This has been confirmed in the present study. Adequate concentrations of micropores in such material could cause an apparent lowering of density and refractive index into the ranges proper to chalcedony or perhaps even to opal (Midgley, 1951; Weymouth and Williamson, 1951).

In the present study two silcretes and two cemented aeolian sands examined casually in thin section appeared to be rich in opal; they failed to yield this mineral when finely ground and subsequently centrifuged with liquids of appropriate density. The same liquids effectively separated powdered synthetic mixtures of opal and quartz, i.e. opal floated and quartz sank.

More detailed examination showed that much opal-like material in the silcretes had a relatively high refractive index and doubtless, as suggested by the results of centrifuging with liquids, a relatively high density. Although the birefringence of the material was low or not apparent it should probably be called chalcedony or quartz.

The terms quartz, chalcedony, opal, and cryptocrystalline silica are used subsequently with the following implications:

Quartz.—Material with the usual optical properties that varied in size down to grains ca. 10 microns across. Aggregates of the latter grains are described as microcrystalline quartz. Some microcrystalline quartz showed a hazy extinction, but as such an extinction does not necessarily indicate a minutely fibrous structure, the material was not classed as chalcedony (Gilbert in Williams, Turner, and Gibert, 1954, p. 358).

Chalcedony.—Demonstrable fibers; their elongation was usually negative but sometimes positive (quartzine).

Opal.—Material of marked negative relief that was isotropic or had a hazy birefringence suggestive of strain rather than of the presence of a cryptocrystalline aggregate.

Cryptocrystalline silica.—Material that was not manifestly fibrous and lacked the marked negative relief of opal. The refractive index was similar to or slightly less than that normally shown by quartz. This silica was isotropic or weakly birefringent, and much of it was probably quartz.

THE SILCRETES

Introduction.—The South African silcretes are related to the underlying rocks in texture and in the nature and characteristics of their "heavy" min-

erals; according to Frankel and Kent (1937) many of them were formed by the silicification of the clay fractions of soils of varying sand content which had developed on these rocks. Inadequate exposures made the petrography of the unsilicified rocks underlying the Australian silcretes a matter for surmise; it was clear, however, that the rocks were siliceous sediments, e.g. sandstones, that contained various proportions of argillaceous material. This material, as in South Africa, had been silicified, but the larger grains of allogenic quartz were still recognizable although now contained in a silicified matrix.

The rocks to be discussed were collected in the northern regions of the State of South Australia, i.e. from localities 70 miles west of, 13 miles south of, and 51 miles west-southwest of Oodnadatta. The last two localities are in the vicinity of Allandale and Arckaringa Homesteads respectively.

Specimens of similar rocks have reached these laboratories from other areas of the Continent, e.g. from the Northern Territory.

Petrographic descriptions appear under the headings: silcrete host rock, syngenetic nodules, colloform structures, replacement veins, cavities and dilation veins, conglomerates.

Silcrete host-rock.—Quartz was the only important allogenic constituent. Rare allogenic feldspar was largely or perhaps entirely potassic. The most frequent residual "heavy" minerals were zircon and rutile; the relatively large worn clastic grains of rutile were readily distinguishable from the swarms of minute authigenic rutile prisms described later.

The allogenic quartz was monocrystalline or appeared as composite grains of which the largest were fragments of sandstone up to 0.5 cm across. It was surrounded by a variable siliceous groundmass of which parts were a microcrystalline mosaic of quartz grains that presented sutured or interdigitated margins to each other. Other parts were cryptocrystalline silica with occasional fibrous chalcedony. Much of the cryptocrystalline silica appeared as turbid areas with or without a feeble birefringence. These areas lacked the distinct negative relief of opal; typical opal was rare.

This siliceous groundmass was essentially authigenic and had replaced the finer-textured and probably more argillaceous portions of the original rock. It had incorporated the smaller grains of allogenic quartz; these were now either interlocked with and indistinguishable from the authigenic quartz around them or, alternatively, had been dissolved and reprecipitated. That solution could have occurred was indicated by the fretting of the residual allogenic quartz by the groundmass. Moreover, the groundmass sometimes corroded this quartz and even penetrated along the grain boundaries in composites or the cracks in individual crystals. Despite these phenomena certain allogenic quartz grains were in optical continuity with regrowth rims of which some were themselves incomplete and corroded. That such rims grew in their present environment and had not been inherited from the parent rock was revealed by their edges; these were interdigitated with the authigenic groundmass and had acquired opaque inclusions from it. Comparable close associations of corroded and regrown quartz grains were recorded by Williamson (1946).

Rhombohedral carbonate occurred very sparsely in the silcretes and was probably authigenic.

Analyses appear in table 1 but probably represent also portions of syngenetic nodules, especially analysis no. 1.

Syngenetic nodules.—The silcrete host rock contained numerous structures identical with the "syngenetic nodules" or "pseudopebbles" described by Frankel (1952). These were usually darker than the host rock and were up to at least 0.5 cm across. Frankel showed that the darker color could arise from enrichment in titanium dioxide. Both the "nodules" and the host rock of the present study contained innumerable minute prisms, sometimes twinned, of rutile. Shapeless opaque or almost opaque grains that accompanied the recognizable rutile may or may not have been of the same mineral. Because of this uncertainty the rutile and accompanying grains will be designated "rutile etc."

It is possible that the grains of rutile etc. were finer, more numerous, and better dispersed in some nodules than in the host rock, but not volumetrically more abundant; this would suffice to explain why such nodules were darker in color than their host. Commonly the silica of the nodules, whether allogenic or authigenic, did not differ in texture from that of the host rock; grains of clastic quartz (fig. 1) frequently lay across the boundaries between the two in the manner described by Frankel (1952).

A few nodules contained one or more smaller nodules of darker color.

The boundary between nodule and host rock was made visible by the phenomena (a)—(c), of which (a) and (b) were incompatible:

(a) The nodule was evenly turbid throughout and darker than its matrix.

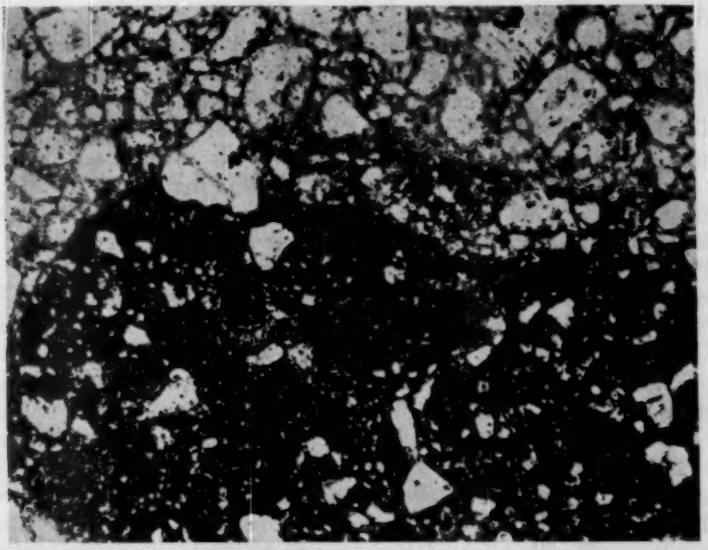

Fig. 1. Dark syngenetic nodule with grain of clastic quartz lying across the boundary with the silcrete host rock (ordinary light; × 33).

(b) At the boundary a zone of enhanced turbidity appeared within the nodule; this zone was ill defined on the inner side. The interior of the nodule sometimes differed little in turbidity from the matrix.

(c) A narrow dark band, which contained grains of rutile etc., marked the boundary and behaved as if it followed a former crack. Some bands had branches which ended blindly or rejoined their parents. Similar dark bands appeared in the silcrete host rock where they followed paths unrelated to the disposition of nodules but again suggestive of former cracks.

Grains of allogenic quartz that lay across the boundary between nodule and host rock broke the apparent continuity of the encircling rutile-rich band; the band was doubtless continuous outside the plane of section.

Certain nodules demarcated according to (c) were no darker than the host rock.

A few nodules indented cavities filled by quartz and chalcedony from which stringers invaded the boundaries between the nodules and the host rock. Such invasion was a further indication that these boundaries had been planes of weakness, i.e. actual or potential cracks. The indentation of the cavities and the presence of the stringers showed that the nodules existed early in the history of the rock; this accords with Frankel's classification of them as "syngenetic" (Frankel, 1952).

Colloform structures.—The term "colloform" was applied by Frankel and Kent (1937) to certain structures in South African silcretes. These were described as titaniferous, cream to cloudy gray in thin section or red-brown from adsorbed iron. They were usually spherical or oval and composed of a series of concentric bands or isolated whorls. Frankel (1952) mentioned cognate titaniferous streaks.

Australian examples could be similarly described with the added comment that abundant rutile was visible in some of them; the mineral formed individual or twinned prisms, usually below 8 microns long, or grouped prisms that sprouted from material that may have been leucoxene.

Frankel (1952) photographed titaniferous streaks which appeared to flow past syngenetic nodules. This disposition of streaks is observable also in some Australian silcrete host rocks, but in addition there were sets of curved but parallel streaks indented in conformity with the margins of syngenetic nodules. Moreover, two phenomena appeared that were not recorded from South Africa. Firstly, some repeated colloform bands, in the host rocks of chalcedony-filled cavities, echoed the outlines of these cavities (fig. 2). Secondly, certain colloform structures occurred in a cryptocrystalline siliceous groundmass, rather free from allogenic quartz grains, that seemed to have been formed by the crystallization of opal. The structures had the appearance, perhaps deceptive, of having originated by the precipitation of titania on the walls of a series of shrinkage cracks. Where, in addition, they echoed the outline of a chalcedony-filled cavity, the suggestion arose that the cavity was caused by the same shrinkage effect as had produced the hypothetical cracks.

The idea that titania was precipitated in cracks was strengthened by the occurrence of the titaniferous streaks which bounded certain syngenetic nodules; evidence was given earlier that such streaks had followed cracks.

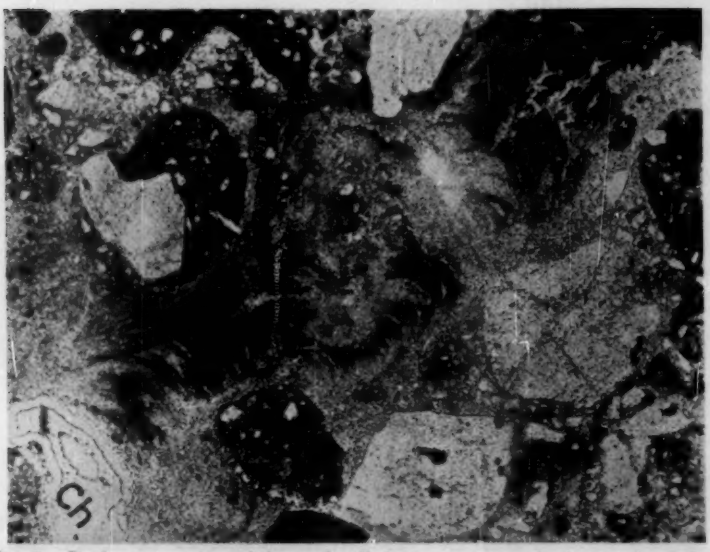

Fig. 2. Dark syngenetic nodules and colloform structures; both contain rutile. Some colloform structures conform with the outline of the chalcedony-filled cavity marked. Ch (ordinary light; × 33).

The precipitation of the titania was presumably in cracks within a siliceous gel. Local movements within this siliceous host, some of which may have accompanied its tendency to crystallize, could annul the former cracks and spread the titania until its disposition was banded rather than linear. (The angle between the plane of section and that of the former crack will also modify the apparent breadth of the titaniferous streak.)

Frankel and Kent (1937) classified the colloform structures as a type of Liesegang phenomenon. This classification is not necessarily invalidated even if it is assumed that systems of cracks were implicated; periodic crystallization could still have been involved.

Replacement veins.—These are illustrated by three examples:

(1) In slice an area of paler was in contact with an area of darker silcrete; the latter area could have been part of an unusually large syngenetic nodule. The paler silcrete contained fewer opaque or almost opaque inclusions of rutile etc. In the darker silcrete, however, such inclusions were not only more numerous but finer grained and better dispersed. Further, the groundmass between the allogenic quartz grains contained considerable proportions of cryptocrystalline or isotropic silica. The groundmass in the paler silcrete was often coarser. It was an aggregate of quartz individuals, commonly microcrystalline, with some cryptocrystalline silica, and flexuous veins of it appeared to enter the darker silcrete (fig. 3). Here their lateral boundaries were either indefinite or marked by discontinuous lines of ferruginous material, possibly formed by segregation. These veins were probably the paths along which permeating solutions had effected some recrystallization. The solutions may have introduced silica but this is not a necessary hypothesis.

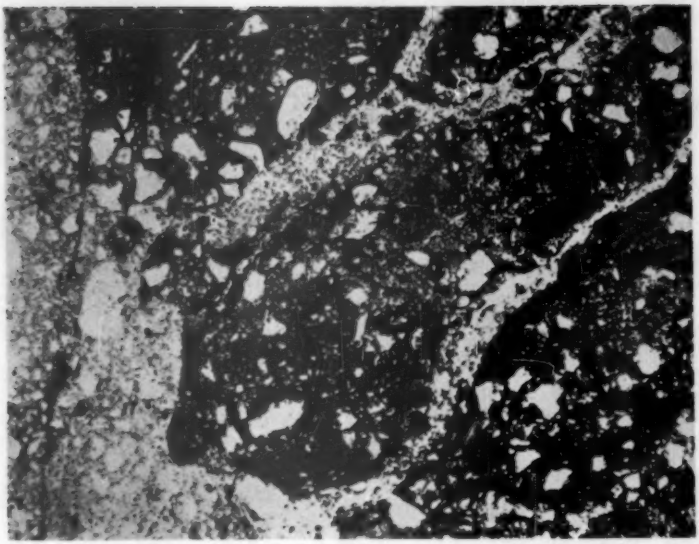

Fig. 3. Dark silcrete apparently veined by matrix of pale silcrete. The "veins" are believed to be recrystallization paths (ordinary light; × 33).

Presumably the paler silcrete had been extensively permeated but the darker rock had been affected only along certain channels.

(2) Individual grains of clastic quartz were penetrated by the surrounding authigenic association of microcrystalline quartz, cryptocrystalline silica and minute rutile prisms. This penetration produced veins of which some were blind and had blunt terminations; in structure, but not in composition, they were reminiscent of certain veins ascribed by Holmes (1936) to metasomatic replacement of quartz. Other veins traversed the clastic quartz grains completely, and many were impossible to classify with confidence as having been produced by dilation or replacement (Goodspeed, 1940; Wells and Bishop, 1954).

(3) Fragments of silcrete, up to 2 cm across and usually angular, appeared in a milky or gray-blue matrix of waxy luster. In slice this matrix was a mosaic of small interlocked areas which, although not manifestly fibrous, had a hazy extinction; in accord with its appearance in hand specimen it is subsequently called chalcedony. If the matrix were absent, most of the silcrete fragments could not be fitted together. Thus on casual examination the rock appeared to be silcrete rubble cemented subsequently by chalcedony; in thin section, however, some of the chalcedony was seen to have replaced silcrete (fig. 4). The silcrete was vaguely or sharply demarcated from the chalcedony but, even where the demarcation was apparently sharp, grains of allogenic quartz frequently straddled the boundary. Very significant were a few areas of silcrete which had already begun to "fade" into the chalcedonic matrix. Grains of allogenic quartz were preserved in the chalcedony on the margins of such areas, while residual bands of authigenic rutile prisms continued from

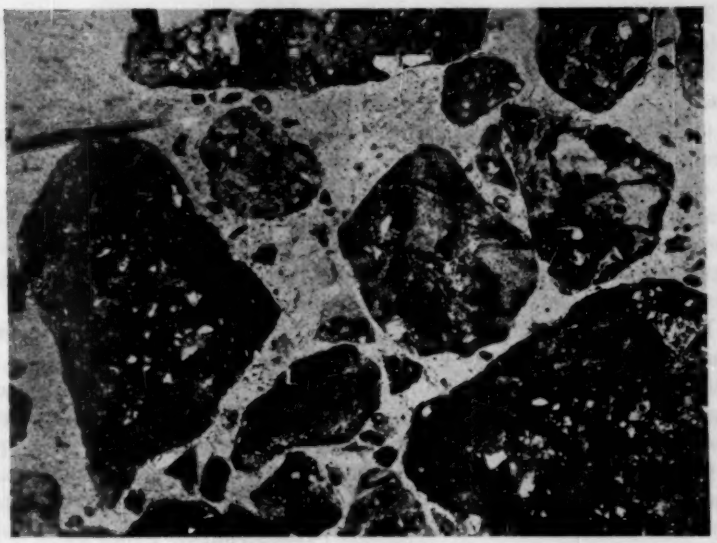

Fig. 4. Breccia of silcrete fragments in chalcedony (ordinary light; × 19).

the silcrete into the matrix. Clusters of quartz grains elsewhere in the chalcedony had presumably survived from silcrete fragments that had been more thoroughly "digested."

Part of the chalcedony was probably formed in place by the solution and subsequent crystallization of silica from the silcrete, while part was introduced; introduced material could have originated by solution of silcrete elsewhere in the immediate vicinity. It would be difficult, however, to refute the suggestion that all the chalcedony had been formed by recrystallization in place. The problem resembled that presented by the apparent replacement of tracts of country rock during granitization. It was made particularly acute because the silcrete and chalcedony were of essentially the same chemical compositions. This virtual identity of composition is true of all the replacement veins and host rocks mentioned under (1)—(3). Thus the term "replacement vein" is used with an extension of its usual significance, i.e. the replacement was of one texture by another, rather than of one chemical compound by another; "recrystallization path" might be an acceptable alternative.

Cavities and dilation veins.—Although cavities and dilation veins were frequently connected, some cavities did not associate with veins, at least in the plane of section. Both cavities and veins contained one or more of the minerals opal, cryptocrystalline silica, fibrous chalcedony, and quartz. Some cryptocrystalline silica may have developed from opal, while there was strong evidence of the development of quartz from chalcedony (Friedman, 1954); fibrous chalcedony was sometimes attached to well crystallized quartz by an intermediate zone in which the fibrous structure had been obliterated only partially. The phenomenon was reminiscent of the evolution of discrete crysals of rhombohedral carbonate from spherulites (Williamson, 1946).

The following criteria were held to indicate when silica minerals had developed in open spaces and not by replacement:

(a) Cavities were incompletely filled, e.g. lined with botryoidal chalcedony.

(b) Cavities and veins were composite and contained, e.g., quartz centrally and chalcedony or cryptocrystalline silica laterally.

(c) Their walls were coated with red-brown iron oxide which sometimes appeared also as independent stringers in the silcrete host rock. The iron oxide seemed to have been deposited before the silica of the veins and cavities; this supposition was strengthened by the fact that some allogenic quartz grains were stained by iron oxide despite their complete inclusion in veins.

The following example was apparently a dilation vein, but detailed examination made its status less clear:

A syngentic nodule, ca. 0.5 mm across in slice, had been faulted and its two portions offset; they showed a separation of 50 microns in the plane of the section, and the space between them was occupied by a vein of quartz and cryptocrystalline silica continuous with the authigenic groundmass of the surrounding silcrete host rock. The two walls of the vein, as seen in the plane of section, fitted neither in length nor shape, but this did not constitute evidence of replacement (Wells and Bishop, 1954). The occurrence of offset suggested a dilation vein (Goodspeed, 1940). However, the continuity of the vein filling with the authigenic groundmass made it seem probable that some replacement had occurred but along a fault crack that provided an easy access for solutions (Webb, 1947).

In authentic replacive structures the spaces in which silica minerals were forming may have been enlarged contemporaneously. This could occur because of the pressure exerted by growing crystals or because of the shrinkage of the matrix. Such shrinkage may, for instance, have accompanied the contraction of a gel. Moreover, it has recently been shown by Roberts and Vivian (private communication) that cracks in mortar may be enlarged by solutions of certain metallic ions; this is thought to depend on adsorption of the ions with consequent mutual repulsion of the two surfaces of the crack, particularly at the apex. It is not yet known whether a similar effect operates in rocks.

Conglomerates.—Two examples are cited because of their analogies with South African occurrences:

(1) Subangular to rounded fragments of a whitish silcrete, up to several inches across, were contained in a second ocher-colored silcrete. Frankel and Kent (1937) and Frankel (1952) have ascribed similar rocks to the operation of repeated periods of silicification; the older silcrete disintegrated during a period of "inter-silicification," and the resulting rubble was cemented when silicification recurred.

(2) This resembled the specimen just described, but the matrix of the fragmental silcrete was not a second silcrete but an association of reddish-brown iron oxide and authigenic quartz. This matrix contained grains of detrital quartz and frequent pores and small cavities. The iron oxide tended to coat the fragments of silcrete and the allogenic quartz grains. Thus it was presumably older than the authigenic quartz. Assemblages of crudely formed

parallel prisms were frequent in this quartz. They were broad and positively elongated and appeared to have developed from fibrous chalcedony.

Fragmental silcrete in a ferruginous matrix is found also in South Africa where it may associate with "ferricrete," i.e. lateritic material (Frankel and Kent, 1937).

CEMENTED AEOLIAN SANDS

The specimens came from two localities in the north of Western Australia, i.e. Minari, near Broome, and Cape Levêque; similar rocks could probably be collected in many other Australian localities. Most of the specimens were so alike that separate description of material from the two localities is not merited.

In material from Minari and Cape Levêque the siliceous matrix occupied 37.2 and 33.1 percent by volume respectively. In these quartzites the detrital quartz was markedly rounded, evenly sized, and often 0.5-0.75 mm diameter; the "heavy" detrital minerals, commonly tourmaline and opaque iron oxides, were markedly rounded also. The universal rounding of the clastic grains, irrespective of their density, and the rather evenly sized quartz grains, showed that the rocks were cemented aeolian sands.

Most quartz grains were monocrystalline but a few were composite; of the latter a small minority were micro- or cryptocrystalline or contained fibrous chalcedony. The quartz grains were slightly fretted against the siliceous matrix, and, as this invaded cracks in some of them, the fretting seemed better attributed to corrosion in place than to the effects of impact during aeolian transport. The volume of the matrix was no greater than that of the void spaces in some of the air-deposited sands prepared experimentally by Kolbuszewski (1948; 1953). Thus it was not possible to determine if crystallization or other effects in the matrix had tended to force the detrital quartz grains apart.

Despite the evidence of corrosion and invasion by the matrix, some quartz grains were welded together along narrow necks, while others had their former rounded outlines modified by irregularly bounded optically continuous rims of secondary quartz (Williamson, 1946).

Chemical analyses (table 1, nos. 4 and 5) showed that the matrix was highly siliceous. It was somewhat turbid in ordinary light and isotropic to just discernibly cryptocrystalline between crossed nicols; the turbidity may have been caused by minute pores or solid inclusions (Weymouth and Williamson, 1951). No opal was obtained when the finely powdered rocks were centrifuged with liquids of appropriate densities. X-ray oscillation photographs of chips of the matrix revealed only very fine-grained quartz.

The apparent refractive index of the matrix, found by immersion, varied slightly. It was commonly 1.546-1.549; the higher values were for slightly ferruginous parts of the matrix of specimens from Cape Levêque. Certain minute chips obtained by grinding quartzite from Minari were composite and included portions of both detrital quartz and siliceous matrix. The matrix was isotropic and had a refractive index of 1.547, which was also the mean

refractive index of the adherent quartz. Winchell (1924) described the same phenomenon.

The matrix in some specimens from Minari surrounded not only rounded quartz grains but small quartz chips, often ca. 50 microns across, with a few fragments of tourmaline not markedly rounded. Occasionally the chips were much larger, e.g. 0.3 mm long, and the rounded grains sparser. The origin of these rocks was made more obscure by the lack of data on their field occurrence.

At Cape Levêque a "porcellanite" was collected, but its field relations to the cemented sands are unknown. It contained quartz grains of different sizes, usually below 0.1 mm across, of which some were rounded but many were angular or subangular; these allogenic grains were cemented by a small volume of cryptocrystalline silica. There were a few flakes of pale mica. The rock appeared to be a variety of silcrete less completely silicified than the examples previously discussed. It was not iron-stained, but cemented sands from the same area frequently were; this accounted for the relatively large percentage of iron shown in analysis no. 5 (table 1). The stained rocks contained numerous tiny authigenic aggregates of a red-brown iron oxide; these were often ca. 8 microns across and occurred mostly in the siliceous cement.

SUB-BASALTIC QUARTZITES

Introduction.—The petrography of the specimens is described only briefly, as most localities merited more detailed field study than was possible in the time available for collecting; such study has, however, been made at Badgery's Lookdown by Waterhouse and Browne (1929) who regarded the quartzite at that locality as a sand mantle, produced by the subaerial weathering of an underlying sandstone, subsequently silicified by a basaltic extrusion. The quartzites at the remaining localities mentioned below were probably formed in an essentially similar way, although fluviatile or lacustrine sands may have been involved in some instances.

Of the localities listed (a)—(e) are in New South Wales and (f)—(g) in Victoria.

Petrography.—(a) *Badgery's Lookdown (Tallong, near Goulburn).*—Grains of detrital quartz, below 2 mm across, usually angular or subangular and commonly monocrystalline, were very variable in size and frequently interlocked in a manner suggestive of thorough recrystallization. In some slices spaces occurred between groups of interlocked grains and these were filled with clear authigenic silica, which was an association of crenulated quartz units and cryptocrystalline material not obviously fibrous. Opal appeared sporadically, usually as water-clear material that surrounded isolated grains of detrital quartz and traversed them along cracks.

Some quartz grains had corroded surfaces with ferruginous coatings; this was reminiscent of certain phenomena in injection provinces ascribed to the passage of residual solutions through the affected rocks (Read, 1927; Williamson, 1935). Despite the corrosion of some quartz grains, others were optically continuous with rims of secondary quartz that had grown beyond the ferruginous coatings.

In parts of the quartzite, clouds of minute dark grains associated with the authigenic silica. These grains were partly and perhaps entirely of rutile and tended to concentrate against the detrital quartz grains and also to appear in the corrosion pits on their surfaces (table 1, analysis no. 6; table 2, analyses nos. 7-11).

(b) *Bolairo (Cooma-Adaminaby road).*—Angular fragments of quartz, of which some were relatively large and up to 4 mm across, were occasionally concave to the matrix. They were monocrystalline or composite and could have been derived from granitic rocks or quartz veins. The matrix, which entered cracks in certain quartz fragments, was a mosaic of quartz grains that contained clusters of minute rutile prisms. These rutile prisms were concentrated asymmetrically against some of the quartz fragments. There were a few shreds of pale mica (table 2; analysis no. 12).

(c) *Bombala (in the extreme S.E. of New South Wales).*—Rounded to angular quartz fragments usually below 0.4 mm across appeared in a mosaic of crenulated quartz grains of various sizes; the smaller grains tended to be the most markedly crenulated. Certain quartz fragments were themselves crenulated against this mosaic, but the crenulation sometimes involved not the edges of the actual fragments but the peripheries of optically continuous rims of secondary quartz. Clouds of minute dark grains appeared in the finer-textured portions of the rock, and at least some of these were rutile (table 2; analysis no. 13).

(d) *Dalton (near Gunning, between Goulburn and Yass).*—Petrographically the specimens did not differ markedly from those from Bombala except that the clouds of dark grains, which included rutile, tended to occur in patches which contained groups of clastic quartz fragments. The dark grains were outside these fragments and associated with an interstitial quartz mosaic (table 2; analysis no. 14).

(e) *Myalla road (ca. 4 miles from Cooma).*—In slice were seen isolated or grouped grains of allogenic quartz. Some of these grains were slightly in excess of 0.4 mm across. The contacts between quartz grains within the groups were not markedly crenulated. Strings and patches of small mutually crenulated quartz grains were seen between the allogenic individuals or groups, towards which they tended to show sutured boundaries. The rock seemed to have suffered a marked recrystallization that had been resisted by the larger allogenic quartz grains. The small crenulated quartz grains may or may not have represented added material (table 2; analysis no. 15).

(f) *Broadmeadows (near Melbourne).*—Relatively large angular monocrystalline or composite quartz grains, sometimes over 3 mm long, appeared to have been granitic debris; granodioritic rocks were exposed nearby (Gaskin, 1944). Some grains were concave to the matrix. The matrix, which invaded certain grains along cracks, was a mosaic of quartz crystals with sutured boundaries; it contained numerous relatively small angular clastic fragments of quartz and many clouds of tiny rutile prisms. These prisms, as in the rock from Bolairo, were sometimes concentrated asymmetrically against the larger fragments of clastic quartz. The clastic quartz grains had been

corroded, and the corrosion appeared to be more intense near the concentrations of rutile prisms.

(g) *Cave Hill* (*Lilydale, near Melbourne*).—According to Gill (1942) the rocks to be described were originally Tertiary fluviatile deposits. There was considerable variation among the specimens collected.

A sandstone contained quartz grains, mostly less than 0.6 mm across in section, of which many were in mutual contact or separated by a sericitic cement with some associated cryptocrystalline silica. Some grains, however, were separated by chalcedony which radiated outwards from their surfaces as fibrous aggregates; the grain surfaces beneath the chalcedony appeared to have been corroded. The radiating fibers were commonly elongated negatively but sometimes continued outwards without change of direction into a band of positively elongated fibers (quartzine). Around some grains only the band of quartzine appeared.

The chalcedonic fringes radiated outwards from adjacent quartz grains until they ended either in mutual contact or as the walls of unfilled pores.

A few quartz grains were optically continuous with rims of non-fibrous secondary quartz. (table 2; analyses nos. 16-17).

DISCUSSION OF CHEMICAL ANALYSES

Most analyses in tables 1 and 2 were made in connection with a survey of raw materials potentially useful for the manufacture of silica bricks. Thus, for technical reasons, some samples (table 2) were calcined before analysis. Comparison of an uncalcined and a calcined sample from the same specimen (table 1, no. 6 and table 2, no. 7) showed, however, that the titania/silica ratio was not manifestly affected by calcination; this is of significance in later discussion.

The rocks were all highly siliceous; probably none contained less than 95 percent of silica before calcination. In only some, however, was the TiO_2/SiO_2 ratio relatively high. If this ratio is calculated from the percentage weights, and values in excess of 0.01 are arbitrarily regarded as high, the three Australian silcretes nos. 1-3 have high ratios. Comparably high ratios characterize South African silcretes. Sixteen analyses, i.e. most or all of those both available and reliable, taken from the literature (Frankel and Kent, 1937; Frankel, 1952) or made in this laboratory (table 2, nos. 18-21) had ratios in excess of 0.01, ie., 0.012, 0.014, 0.014, 0.015, 0.015, 0.016, 0.016, 0.017, 0.017, 0.020, 0.021, 0.023, 0.023, 0.027, 0.029, and 0.030. Such ratios are not characteristic of siliceous sediments and some of them would be high even for argillaceous materials. Thus 0.003 and 0.005 are possible averages for sandstones, and 0.013 and 0.008 for shales (Clarke, 1924; composite analyses of 253 and 371 sandstones and of 51 and 27 shales, i.e. analyses F and G, p. 547, and A and B, p. 552).

The titania in the silcretes is present partly, and probably largely, as the rutile etc. which is especially concentrated in the colloform structures and syngenetic nodules. Concentration in these structures and nodules showed that the titaniferous minerals were not merely residual crystals inherited from the rocks converted to silcretes. On the contrary, titanium had been deposited

from ionic or colloidal solutions by a concentration process comparable with that undergone by silicon.

The titania/silica ratios in certain sub-basaltic quartzites were relatively high (nos. 11 and 12). They may or may not reflect similar ratios that existed in the sediments prior to conversion into quartzites. However, the tendency for rutile to concentrate in clouds of prisms, sometimes appearing asymmetrically against fragments of clastic quartz, suggested that titanium may have entered in solution from the overlying basalts.

THE ORIGINS OF THE SILICIFYING SOLUTIONS

Silcretes.—According to Frankel and Kent (1937) the South African silcretes were derived from the subsoils of residual deposits through the agency of upwardly migrating solutions of colloidal silica; these solutions were formed during the quiescent leaching of the underlying rocks and became coagulated when they met downwardly migrating solutions rich in sodium chloride. Such an hypothesis of the coagulation of silica by electrolytes, whatever its merits, could be readily applied in Australia where the clays associated with the "duricrust" frequently contain sodium chloride, gypsum, or alunite.

Mountain (1952) offered chemical evidence which suggested that the silcrete in a clay zone derived from the Dwyka tillite represented the silica lost during the formation of the clay. As silcretes occur on or in such clay zones, at least part of the silica for their formation would appear to have migrated upwards. Where, however, silcretes rest on and pass gradationally into subadjacent siliceous rock, e.g. the Witteberg quartzite mentioned by Mountain (1952), some at least of the silica presumably traveled downwards and originated by solution of the siliceous rock especially near the exposed surfaces.

Any valid hypothesis of the mechanism of silcrete formation applicable in South Africa will probably apply also to Australian occurrences. The latter, however, require more detailed study before this can be asserted with confidence.

Cemented aeolian sands.—The rocks from Minari and Cape Levêque present genetic problems the elucidation of which must await, among other factors, adequate field evidence. Siliceous solutions of unknown origin entered unconsolidated sand from an unknown direction. An upward movement caused by capillarity is not impossible despite the relatively high porosity probably shown by the uncemented sands (Baver, 1948, p. 222 ff); alternatively the solutions may have drained downwards into the sands if the contemporaneous topography permitted it (Bassett, 1954). A further possibility is that of silicification in the beds of intermittent lakes.

Unlike the solutions involved in the formation of the silcretes, those which affected the aeolian sands appear to have been poor in titanium (table 1, nos. 4 and 5).

Sub-basaltic quartzites.—Those without opal or chalcedony may have been recrystallized by a type of contact metamorphism in which the role of solutions was unimportant. In common with quartzites which contained these

minerals they had, however, a proportion of fine quartz grains with crenulated or sutured junctions. Waterhouse and Browne (1929), in their account of the quartzites of Badgery's Lookdown, wrote of this quartz as probably deposited from solution; although crenulated and sutured junctions may be shown by minerals in hornfelses, strength is lent to these investigators' suggestion by the presence of opal and chalcedony, indicative of the former existence of siliceous solutions. At least five hypotheses could be advanced in explanation of the origin of such solutions. Numbers (1) and (3) were discussed by Waterhouse and Browne (1929) who preferred (3).

(1) Pore water became heated and was thus a more effective solvent for silica already present within the rock.

(2) Water or solutions, presumably hot, came from the basalt and served as the solvent, or (3) introduced silica.

(4) The silicifying solutions originated during the weathering of the solidified basalt. This was suggested by Frankel and Kent (1937) for an Australian occurrence like that at Badgery's Lookdown.

(5) The same investigators preferred the alternative that the solutions were unconnected with the basalt and had migrated upwards, i.e. the resulting rock was cognate with a silcrete. This is not likely to be acceptable to geologists with field experience in Australia.

The first four hypothetical mechanisms are not mutually incompatible, and more than one may have contributed to the silicification process.

SUGGESTED FURTHER INVESTIGATIONS

One purpose of the present reconnaissance was to direct attention to the rocks described in the hope that others would investigate them more thoroughly. Geochemical studies of the South African silcretes, especially to elucidate the fate of trace elements during silicification, could be one enquiry. Similar study would be desirable in Australia but only after complete profiles of silcretes and associated rocks had been investigated in the thorough manner demonstrated by Frankel and Kent (1937).

Geochemical techniques applied to the sub-basaltic quartzites would probably reveal if elements had immigrated into them from the overlying basalts and thus could give information as to the sources of the silicifying solutions.

SUMMARY

(1) Silicified surface rocks included silcretes like those of South Africa, aeolian sands, and material indurated beneath basalt and subsequently exposed by removal of the igneous cover (sub-basaltic quartzites).

(2) The silcretes, although collected in the northern part of the State of South Australia, are believed to occur widely in Australia.

(3) They were highly siliceous but enriched also in titania which was especially concentrated as rutile in "colloform structures" and "syngenetic nodules." South African silcretes have all these characteristics.

(4) The Australian silcretes are believed to represent siliceous sediments, e.g. sandstones, that contained various proportions of argillaceous material. The latter material, as in South Africa, had been silicified.

(5) Siliceous veins in silcrete were ascribed to dilation or replacement. The replacement veins may be alternatively called recrystallization paths as along them the grain size of silica minerals appeared to have altered in response to percolating solutions. The replacement was of one texture by another, rather than of one compound by another. The phenomenon is reminiscent of those in "permeation gneisses."

(6) Quartzites from localities in the north of the State of Western Australia were cemented aeolian sands. These also are believed to occur widely in Australia.

(7) Selected sub-basaltic quartzites from New South Wales and Victoria were described. At least some of them appeared to have developed from unconsolidated siliceous material, e.g. sands. Possible genetic mechanisms are outlined.

(8) The problem of identifying opal, chalcedony, and quartz became very apparent during the present study. This, and the nomenclature of the three minerals, are discussed.

(9) Opal, chalcedony, and authigenic quartz tended to corrode and invade clastic quartz in all types of surface rock.

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COMMONWEALTH SCIENTIFIC AND INDUSTRIAL RESEARCH ORGANIZATION
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TABLE 1

Analysis of Australian Silicified Rocks

(Samples dried at 105°C)

No.	1	2	3	4	5	6
SiO_2	95.35	96.22	96.44	98.82	98.29	97.27
TiO_2	1.26	1.71	1.80	Tr.	Tr.	0.20
Al_2O_3	0.36	0.73	0.96	0.00	0.19	0.10
*Fe_2O_3	1.81	0.24	0.28	0.29	1.00	1.90
CaO	0.33	0.40	0.31	0.07	0.08	0.20
MgO	0.09	Tr.	Tr.	0.01	0.01	0.05
Na_2O	nil	n.d.	n.d.	n.d.	n.d.	n.d.
K_2O	0.02	n.d.	n.d.	n.d.	n.d.	n.d.
Loss on ignition	0.73	0.77	0.58	0.24	0.41	0.55
Total	99.95	100.07	100.37	99.43	99.98	100.27
TiO_2/SiO_2	0.013	0.018	0.019	0.000	0.000	0.002

* State of oxidation not determined.

1 Silcrete (composite sample), Allandale Homestead (13 miles south of Oodnadatta).

2 Silcrete, Allandale Homestead (13 miles south of Oodnadatta).

3 Silcrete, Arckaringa Homestead (7 miles west of Oodnadatta).

4 Cemented aeolian sand, Minari.

5 Cemented aeolian sand, Cape Levêque.

6 Sub-basaltic quartzite, Badgery's Lookdown (same sample as no. 7, table 2).

Analyst: Miss B. C. Terrell.

TABLE 2

Analyses of Australian Sub-basaltic Quartzites and South African Silcretes
(Samples calcinated at 1425°C for 4 hours)

No.	7	8	9	10	11	12	13
SiO_2	97.15	96.74	99.14	99.03	97.84	98.27	n.d.
TiO_2	0.20	0.14	0.18	0.32	1.27	1.08	0.24
Al_2O_3	0.10	0.22	0.28	0.22	0.60	0.39	0.13
*Fe_2O_3	1.90	2.18	0.16	0.17	0.21	0.27	0.10
CaO	0.39	0.08	0.13	0.10	0.06	0.13	n.d.
MgO	0.05	Tr.	Tr.	Tr.	Tr.	Tr.	n.d.
Total	99.79	99.36	99.89	99.84	99.98	100.14	n.d.
TiO_2/SiO_2	0.002	0.002	0.002	0.003	0.013	0.011	n.d.

No.	14	15	16	17	18	19	20	21
SiO_2	99.07	99.02	n.d.	n.d.	95.92	97.43	97.91	96.93
TiO_2	0.53	0.29	0.11	0.22	2.82	1.35	1.32	1.53
Al_2O_3	0.14	0.32	0.14	1.98	0.84	1.78	0.55	0.90
*Fe_2O_3	0.18	0.18	0.06	0.28	0.36	0.21	0.37	0.27
CaO	0.08	0.14	n.d.	n.d.	0.06	0.11	0.10	0.09
MgO	Tr.	Tr.	n.d.	n.d.	nil	Tr.	Tr.	Tr.
Total	100.00	99.95	n.d.	n.d.	100.00	100.88	100.25	99.72
TiO_2/SiO_2	0.005	0.003	n.d.	n.d.	0.029	0.014	0.014	0.016

* State of oxidation not determined.

7 Sub-basaltic quartzite, Badgery's Lookdown (same sample as no. 6, table 1).

8 Sub-basaltic quartzite, Badgery's Lookdown

9 Sub-basaltic quartzite, Badgery's Lookdown

10 Sub-basaltic quartzite, Badgery's Lookdown

11 Sub-basaltic quartzite, Badgery's Lookdown (nos. 7-11 represent hand specimens differing in texture and color).

12 Sub-basaltic quartzite, Bolairo.

13 Sub-basaltic quartzite, Bombala (silica percentage over 99).

14 Sub-basaltic quartzite, Dalton.

15 Sub-basaltic quartzite, Myalla Road.

16 Sub-basaltic quartzite, Cave Hill (silica percentage ca. 99.5).

17 Sub-basaltic quartzite, Cave Hill; Hand specimen of different appearance from no. 16 (silica percentage ca. 96.0).

18 Silcrete near Albertinia, Cape Province, South Africa. Replaced clay (Karnemelksvlei).

19 Silcrete near Albertinia, Cape Province, South Africa. Replaced clay (Middledrift).

20 Silcrete near Albertinia, Cape Province, South Africa. Nodular variety.

21 Silcrete near Albertinia, Cape Province, South Africa. Nodular variety.

Analyst: Miss B. C. Terrell.

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CORDIERITE-INDIALITE RELATIONS

AKIHO MIYASHIRO

ABSTRACT. It was recently found that minerals in structurally intermediate states between indialite (hexagonal) and common cordierite (pseudohexagonal) exist in nature as well as in artificial products. The structurally intermediate minerals show distinctive X-ray diffraction patterns and optical properties. This finding has led to a serious revision of our knowledge of the minerals of the cordierite group.

On heating at high temperatures, cordierite is transformed into indialite continuously through all gradations. Probably the transformation is of order-disorder type. The stability relations of the minerals of the cordierite group have been determined. The existence of high and low modifications in indialite as well as in cordierites is discussed.

A revised structural classification of the minerals of the cordierite group is presented in the concluding section of this paper.

INTRODUCTION

Indialite is a hexagonal mineral, polymorphic with cordierite (Miyashiro and Iiyama, 1954; Miyashiro and others, 1955). Miyashiro and others (1955, p. 204) wrote: "The existence of crystals in some structurally intermediate state between indialites and cordierites is not impossible, as the crystal structures of cordierites are derived by slight deformations from those of indialites. However, we have not found such an example."

In the summer of 1955, Prof. M. Sambonsugi of Hukusima University sent a mineral from a granitic pegmatite at Sugama in Hukusima Prefecture to Prof. H. Kuno of Tokyo University for identification. The mineral was given to me for examination and I came to the conclusion that it is structurally intermediate between indialite and common cordierite. This finding prompted me to resume a study of cordierite and indialite with special attention to the intermediate states.

Soon I found that such structurally intermediate minerals are not rare either in nature or in artificial products, and it was only from lack of due attention that the intermediate minerals had not been found. All gradations appear to exist between indialite and common cordierite. Structurally intermediate minerals show distinctive X-ray diffraction pattern and optical properties. Heating experiments have clarified the stability relations of the minerals of the cordierite group. The relations between indialite and cordierite resemble those between sanidine and microcline discussed by Goldsmith and Laves (1954a, 1954b).

Though Miyashiro and others (1955) discussed the existence of high and low modifications in cordierite as well as in indialite, the problem is discussed from this new point of view in this paper.

DISTORTION INDEX

Indialite is iso-structural with beryl, whereas the structure of cordierite may be derived by slight distortion from that of indialite. X-ray diffractograms for indialite and cordierite show a general similarity. The pattern in the range of 2θ between 29° and 30° for CuKα radiation is the most distinctive. In this range, at least three peaks (possibly four or five peaks) appear in cordierite, while all these peaks are united to form a single peak in indialite, as shown in figure 1. The three highest peaks of the cordierite pattern between

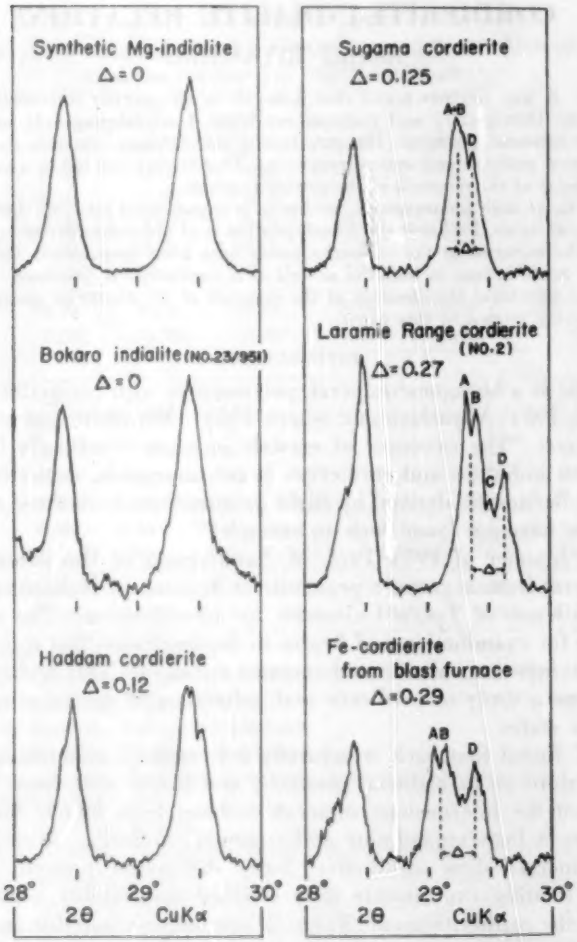

Fig. 1. X-ray diffractograms of indialites and cordierites.

29° and 30° are designated as *A*, *B*, and *D* from the lower angle to the higher. A fourth peak which may appear between the peaks *B* and *D* is designated as *C*. (Probably the peaks *A*, *B*, and *D* represent the reflections (511), (421), and (131), respectively, of CuKα_1 radiation. The peak *C* is probably due to the reflection (421) of CuKα_2 radiation. The corresponding single peak of indialite represents the (12$\bar{3}$1) reflection.)

The spaces between *A*, *B*, and *D* increase with the degree of distortion of the cordierite structure. In indialite all these spaces are equal to zero. Therefore, these spaces or some quantities related to them can be used as an index to represent the degree of distortion. It has been found that the following value Δ is the most convenient as an index of the degree of distortion:

$$\Delta = 2\theta_D - \frac{2\theta_A + 2\theta_B}{2}$$

where $2\theta_A$, $2\theta_B$, and $2\theta_D$ indicate the deviation angles, in degrees, of the peaks *A*, *B*, and *D*, respectively, for CuKα_1 radiation. This value, called the *distortion index* in this paper, is equal to zero for indialite, and increases continuously as the distortion increases. The maximum value of the distortion index has been found to be of the order of 0.29-0.31 for any Fe^{+2}/Mg ratio of the mineral.

Cordierite showing the maximum degree of distortion is called *perdistortional cordierite* in this paper, whereas the minerals structurally intermediate between indialite and perdistortional cordierite are called *subdistortional cordierite*. Thus the distortion index of the minerals of the cordierite group is as follows:

Cordierite	Perdistortional cordierite	$\Delta \geqslant 0.29$
	Subdistortional cordierite	$0.29 > \Delta > 0$
Indialite		$\Delta = 0$

In some subdistortional cordierites (for example, in the Sugama cordierite of fig. 1), the peaks *A* and *B* cannot be resolved from each other, thus making an apparently single peak. In such cases, the reading of the united peak of *A* and *B* may be considered to represent $(2\theta_A + 2\theta_B)/2$, because the peaks *A* and *B* usually have similar heights. Then the distortion index of such a mineral is equal to the space between the united peak and the peak *D*. When a sample has a small Δ value and especially when a sample is composed of a mixture of cordierites with small variable Δ values, the resolution of the peaks *A*, *B*, and *D* from one another is bad. Thus, it is not rare that all the peaks *A*, *B*, and *D* are united to form an apparently single peak, broader than a true single peak. The nature of such samples can be inferred from the width of the peak.

The complete indexing of the powder diffraction data of many cordierites with various compositions and Δ values has been performed by Mr. T. Iiyama and published in a separate paper (Iiyama, 1956).

The distortion is probably due to some order-disorder change of the Si and Al atoms in the cordierite-indialite rings Si_5AlO_{18}. Probably the disordered arrangement produces the hexagonal symmetry of indialite, while an ordered arrangement produces the lower symmetry of perdistortional cordierite. Subdistortional cordierite represents an intermediate degree of order. Thus, the distortion index is probably a measure of the degree of order. Rearrangements of Al, Mg, and Fe^{+2} atoms between the rings may accompany the order-disorder transition.

OBSERVED DISTORTION INDICES OF NATURAL CORDIERITES

Introductory Statement

The distortion index and optical properties were measured on various natural cordierites and indialite, as shown in table 1. The possible magnitudes of error are ± 0.01 for Δ, $\pm 5°$ for 2V, and ± 0.003 for β. The $(Fe^{+2}+Mn) \times 100/(Mg+Fe^{+2}+Mn)$ ratios given in the last column were calculated

TABLE 1

The Distortion Indices and Optical Constants of Natural Cordierites and Indialite

Locality	Host rock and mode of occurrence	Δ	2V about X	β_D	$\frac{Fe^{+2}+Mn}{Mg+Fe^{+2}+Mn}\times 100$
Bokaro, India	Fused sediments	0.00-ca. 0.31	0°-ca.90°		
Asama, Japan	Xenolith in andesite	0.30	92°	1.544	
Etna, Italy	Xenolith in 1669 lava	0.29			
Kasyo-to near Taiwan	Crystals in andesite	0.25	88°	1.537	
Seto, Japan	Xenocrysts in granite-porphyry	0.27		1.542	
Laramie Range, Wyoming (no. 1)	Metamorphic deposit	0.27	77°-88°	1.548	About 33.0*
Laramie Range, Wyoming (no. 2)	Metamorphic deposit	0.27	77°	1.541	11.1
Orijärvi, Finland	Metamorphic rock	0.27	89°	1.536	15.9
Sori, Japan	Biotite-cordierite-hornfels	0.27	70°	1.555	
Komagane, Japan	Sillimanite-cordierite-biotite gneiss	0.27			
Micanite, Colo.	Sillimanite-cordierite-biotite gneiss	0.26	88°		
Tanzawa, Japan	Metasomatic cordierite-biotite rock	0.245			
Sarurugawa, Japan	Porphyroblasts in gneiss	0.21	80°	1.550	
New Haven, Conn.	Pegmatite	0.26	72°	1.544	22.6
Mont Bity, Madagascar	Quartz vein	0.26	68°	1.544	22.2
Risör, Norway	Pegmatite (?)	0.25	84°	1.535	
Hitati, Japan	Quartz vein	0.25	68°	1.546	21.9
Guilford, Conn.	Pegmatitized gneiss	0.24	75°	1.546	
Dosi, Japan	Pegmatite	0.23	66°	1.566	76.7**
Micanite, Colo.	Pegmatite	0.13	49°	1.568	
Sugama, Japan	Pegmatite	0.125	55°	1.574	70.3
Haddam, Conn.	Pegmatite	0.12	42°	1.560	33.6

* Estimated from FeO = 7.06% alone.

** The original specimen described by Shibata (1936).

TABLE 2
Chemical Compositions of Cordierites

	I	II	III	IV
SiO_2	48.04	47.09	46.99	44.64
Al_2O_3	33.41	34.86	32.72	29.96
TiO_2	0.06	0.03	0.10	n.d.
Fe_2O_3	0.01	0.11	0.16	2.72
FeO	5.04	2.59	7.12	11.02
MnO	0.20	0.05	0.42	1.86
MgO	10.32	11.92	8.38	3.08
CaO	<0.04	0.04	0.06	0.02
Na_2O	1.12	0.78	0.85	2.05
K_2O	0.30	tr.	0.30	0.75
H_2O_+	1.81	2.21	3.07	3.85
H_2O_-	0.23	0.65	0.05	
Total	100.58	100.33	100.22	99.95
α_D	1.539	1.536	1.551	1.560
β_D	1.544	1.541	1.560	1.574
γ_D	1.547	1.546	1.562	1.578
2V about X	68°	77°	42°	55°
Dispersion (about X)	Weak, v>r	Very weak. v>r	Weak, v>r	Weak, v>r
Δ	0.26	0.27	0.12	0.125

I: Cordierite from Mont Bity, Madagascar. Analyzed by H. Haramura.
II: Cordierite from Laramie Range, Albany County, Wyoming ("Laramie Range cordierite no. 2"). Analyzed by T. Katsura and H. Haramura.
III. Cordierite from Haddam, Connecticut. Analyzed by H. Haramura.
IV: Cordierite from Sugama, Hukusima Prefecture, Japan. Analyzed by an analyst of the Geological Survey of Japan (Sambonsugi, 1956).

from chemical analyses, four of which are shown in table 2. The observed values of Δ range from 0.00 to 0.31.

The Laramie Range cordierite no. 2, which has a very low Fe^{+2}/Mg ratio, shows $\Delta = 0.29$ after heating at 1000°C for 10 minutes. The Asama cordierite shows $\Delta = 0.30$. The Kasyo-to cordierite, heated at 900°C for 60 minutes, shows $\Delta = 0.31$. An artificial cordierite with a very high Fe^{+2}/Mg ratio shows $\Delta = 0.29$, as will be stated later in detail. Thus, we may consider that the maximum value of the distortion index is of the order of 0.29-0.31 for any Fe^{+2}/Mg ratio.

Distortion Index and Optical Angle

There exists a close relation between the distortion index and optical angle as shown in figure 2. Generally speaking, the larger the distortion index, the larger the optical angle about X. The chemical composition (or the high-low transition to be discussed later) has only a very little effect on the optical angle.

Distortion Index and Refractive Indices

There exists also a relation between the distortion index and refractive indices. Figure 3 shows the relation in cordierite from metamorphic rocks,

pegmatites and quartz veins. Two cordierites with the distortion indices of 0.12 and 0.125 have much higher refractive indices than other cordierites with distortion indices of 0.25-0.27 and with nearly same Fe^{+2}/Mg ratios. The smaller the distortion index, the higher the refractive indices, so far as all the other factors remain constant.

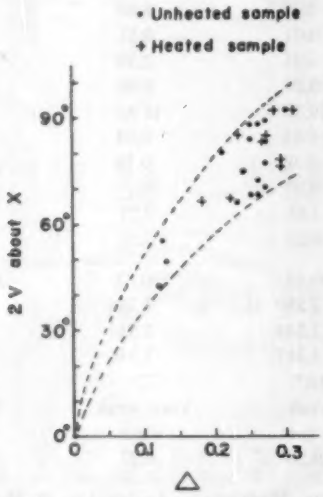

Fig. 2. The relation between the distortion index and optical angle of cordierites and indialite.

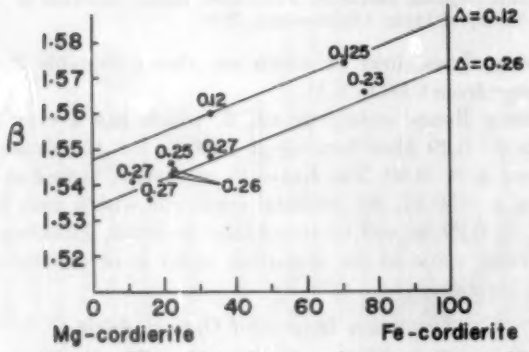

Fig. 3. The relation between the composition and β index of cordierites from metamorphic rocks, pegmatites and quartz veins (i.e. low cordierites). The numerals associated with the points represent the Δ value.

Distortion Index and Modes of Occurrence

Certain samples from volcanic rocks belong to the group of perdistortional cordierite, while the remaining sample belongs to the group of sub-

distortional cordierite. All the samples from metamorphic rocks, pegmatites, and quartz veins belong to the group of subdistortional cordierite.

It is noteworthy that subdistortional cordierite with $\Delta > 0.20$ was found in metamorphic rocks as well as in pegmatites and quartz veins, while subdistortional cordierite with $\Delta < 0.20$ was found characteristically in pegmatites.

INDIALITE AND CORDIERITE FROM BOKARO COALFIELD

In previous papers (Miyashiro and Iiyama, 1954; Miyashiro and others, 1955, p. 193-194), it was stated that the cordierite-like mineral in fused sediments (so-called para-lava) of Bokaro coalfield, India, was actually indialite. Now I have re-examined two specimens of the fused sediments with special attention to the structurally intermediate states. Thus it has been revealed that some of the cordierite-like crystals are true indialite, but others are subdistortional cordierites with variable Δ values.

Specimen No. 23/951

This specimen contains numerous minute colorless crystals embedded in glass. Most of them do not show twinning. According to conoscopic observations, some of them are completely or nearly uniaxial negative. In other crystals, the central portion is uniaxial negative, whereas the marginal portion is biaxial with optical angles about X smaller than 40°. The refractive indices are $\omega_D = 1.539$ and $\varepsilon_D = 1.534$.

The X-ray diffractogram of the powdered rock has only one peak between 29° and 30° for CuKα radiation, as shown in figure 1. The diffraction data have been indexed completely on the basis of a hexagonal unit cell with dimensions $a = 9.812$ Å and $c = 9.351$ Å, as shown in table 5 of the paper by Miyashiro and others (1955). Thus at least most of the crystals in this specimen are considered to be indialite. Probably the uniaxial mineral is indialite and the biaxial one is subdistortional cordierite with very small Δ values.

Specimen No. 23/952

This specimen contains slightly larger crystals, which generally show complicated twinning between crossed nicols, as was described by Venkatesh (1954). The optical angle of the crystals is highly variable, ranging from a small value to nearly 90° about X.

The X-ray diffractogram has an apparently single but too broad peak between 29° and 30° for CuKα radiation, indicating that probably the crystals are subdistortional cordierites with variable Δ values. Probably the average value of Δ is about 0.15. A small number of the crystals may be true indialite or perdistortional cordierite.

Optical Angle and Optical Anomaly

The results of studies, given in the present and preceding parts of this paper, indicate that the optical angle is a fairly good indicator of the structural states in indialite as well as in cordierites. True indialite is completely uniaxial negative or at least nearly so, and the optical angle becomes generally larger with the increasing Δ values, as already shown in figure 2. Thus, the optical anomaly, if any, is not so remarkable in this case.

ARTIFICIAL INDIALITE AND CORDIERITE

The α form of $Mg_2Al_4Si_5O_{18}$, synthesized and named by Rankin and Merwin (1918), is completely or nearly uniaxial negative without twinning. Certainly it is indialite. Several later investigators, however, reported the formation of artificial "cordierite" with varied optical angles, such as 8°, 30°, 50°, and 82°, usually with twinning (Dittler and Köhler, 1938; Shand, 1943; Sugiura, 1951; Yoder, 1952).

I examined a "magnesium-cordierite" with $2V = 8°$, synthesized and described by Sugiura (1951). It is composed of radial aggregates of crystals, resembling twinning. Its unit cell was found to have a hexagonal shape within the limit of experimental error. Thus, it is either true indialite showing slight optical anomaly or subdistortional cordierite with a very small Δ value. It appears probable that twinned artificial "cordierites" with intermediate optical angles, such as 30° and 50°, are subdistortional cordierites with varied Δ values.

Richardson and Rigby (1949) and Richardson, Ball, and Rigsby (1952) described an artificial cordierite with a very high Fe^{+2}/Mg ratio from blast-furnace linings. A chemical analysis of the cordierite-bearing part showed 27.07 percent FeO and 0.68 percent MgO. It is biaxial negative, and certain crystals exhibit complicated twinning. Recently I obtained a sample of the material through the courtesy of Dr. S. O. Agrell of the University of Cambridge. The X-ray diffractogram shows that it is a perdistortional cordierite with $\Delta = 0.29$ (fig. 1).

Thus, it is probable that indialite and subdistortional and perdistortional cordierites all can be formed in artificial processes under various conditions. The conditions related probably comprise not only the temperature of formation but also the rate of crystal growth and the properties of the surrounding materials. As is shown below, indialite is stable at higher temperatures than subdistortional and perdistortional cordierites. Rapid growth of crystals may favor the formation of indialite and subdistortional cordierites with small Δ values even at temperatures where cordierites with larger Δ values are stable.

In a previous paper (Miyashiro and others, 1955, p. 191-192, 194), it was stated that the "twinning", observed between crossed nicols, of artificial "cordierites" and of Bokaro crystals might actually be a kind of apparent optical structure, genetically related to optical anomaly. At that time I regarded those twinned materials as indialite. However, it has now been clarified that they are probably subdistortional cordierite. Thus there remains no reason to doubt that the structure is true twinning. These twins are, in most cases, very complicated in shape, possibly because of variations of the Δ value in different parts of a single crystal.

HEATING EXPERIMENTS OF CORDIERITES

Introductory Statement

Each small sample of natural cordierite was heated at a certain definite temperature in an electric furnace in air and then taken out to cool. The temperature was controlled within $\pm 20°$. The distortion index, optical angle, and refractive indices were measured on the cooled samples.

The rate of structural changes appears to be influenced not only by the heating temperature but also by some unknown factors (for example, possibly by the grain size). Small pieces of cordierite, several millimeters in size, were used as samples in most cases, when available. Certain heated samples were composed of a mixture of cordierites with variable Δ values and optical properties, probably owing to a difference in the rate of structural change in different parts of the samples. In such cases, the estimated average values of Δ, 2V, and/or β are shown in the following tables.

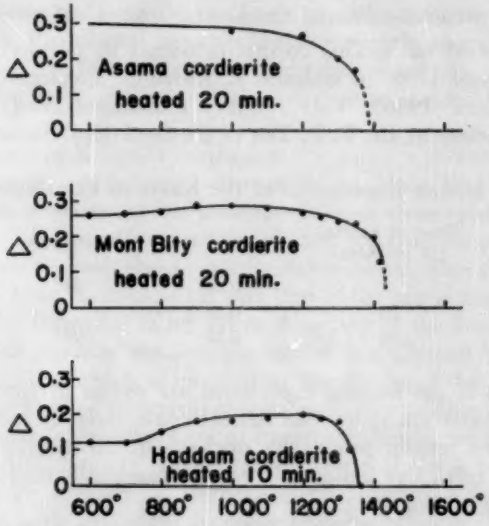

Fig. 4. The changes of the distortion index on heating.

Perdistortional Cordierite

As an example of natural perdistortional cordierite, I used a material from volcano Asama, which lies about 130 kilometers northwest of Tokyo. The cordierite occurs as small grains in xenoliths included in augite-hypersthene andesite. (A detailed petrological and mineralogical description of the cordierite-bearing xenoliths will be made by Mr. S. Aramaki of Tokyo University in the near future.)

TABLE 3

Heating Experiment of the Asama Cordierite

Temperature (°C)	Time of heating (in minutes)	Δ	2V about X	β_D
Room temp.		0.30	92°	1.544
1000	20	0.28	92°	1.547
1200	20	0.265	83°	1.552
1300	20	0.20		
1350	20	0.17		
1400	20	Decomposes		

The results of the heating experiment are summarized in table 3. On heating at high temperatures, the distortion index decreases as shown in figure 4, whereas the optical angle becomes smaller and the β index increases. The color of the mineral becomes darker and brownish. Therefore it is considered that the change in optical angle and refractive indices are due to the gradual structural change towards indialite as indicated by the decrease in Δ, as well as to the oxidation of iron. On heating at 1400°C, the mineral decomposes into liquid and mullite.

Subdistortional Cordierites with $\Delta > 0.20$

Kasyo-to cordierite.—This cordierite occurs as porphyritic individuals, up to 1 centimeter long, in andesite at Kasyo-to, a volcanic island, east of Taiwan (Ichimura, 1936). Very recently I obtained good specimens of it through the courtesy of Dr. T. P. Yen of the Geological Survey of Taiwan.

TABLE 4

Heating Experiment of the Kasyo-to Cordierite

Temperature (°C)	Time of heating (in minutes)	Δ	2V about X	β_D
Room temp.		0.25	88°	1.537
900	20	0.28		1.537
900	60	0.31	92°	1.537

The results of the heating experiment are shown in table 4. On heating at 900°C, the distortion index and optical angle about X increase, while the refractive indices remain practically unchanged. On heating at 1000°C for 20 minutes, the refractive indices still remain practically unchanged. (A more detailed study is in progress.)

Mont Bity cordierite.—This cordierite occurs in a quartz vein cutting through mica schists at Mont Bity, Madagascar. The chemical composition is shown in table 2.

TABLE 5

Heating Experiment of the Mont Bity Cordierite

Temperature (°C)	Time of heating (in minutes)	Δ	2V about X	β_D
Room temp.		0.26	68°	1.544
600	20	0.26		1.538
700	20	0.265		1.538
800	20	0.27		1.536
900	20	0.29	78°	1.536
1000	20	0.29		1.534
1100	20	0.285		1.533
1200	20	0.27	84°	1.533
1250	20	0.26		1.533
1300	20	0.245		1.533
1350	20	0.24		1.533
1400	20	0.20	84°	1.532
1415	20	0.13		

Table 5 and figure 4 show that the Δ value changes on heating and it becomes largest after heating at 900° and 1000°C. This fact may be interpreted as follows: At temperatures below 1000°C, perdistortional cordierite with Δ = about 0.29 is stable. The Mont Bity cordierite is initially in a metastable state with Δ = 0.26. On heating at 700°-1000°C, it gradually changes towards the perdistortional-cordierite state, stable at those temperatures. On heating at 900° and 1000°C for 20 minutes, the stable state is reached. On heating at and above 1100°C, the mineral gradually resumes the subdistortional-cordierite states with smaller Δ values, as such states are stable at those high temperatures. Probably it becomes indialite slightly above 1415°C. (The increase of the Δ value of the Kasyo-to cordierite on heating at 900°C, mentioned before, also may be regarded as a change to the stable state at that temperature.)

In the Asama and Kasyo-to cordierites the β index increases or remains practically unchanged on heating, whereas in the Mont Bity cordierite the β index decreases remarkably on heating. Table 5 shows that the decrease begins at so low a temperature that the change towards the perdistortional-cordierite state, mentioned above, hardly takes place. The Δ value of the sample, heated at 1250°C, is identical with that of the initial material, whereas the β index of the former is much lower than that of the latter. These facts indicate that the decrease of the β index cannot be explained by that kind of structural change alone which is revealed by the change of the Δ value. Being superposed on this kind of structural change, probably another kind also takes place. Thus, as regards the latter kind of structural change, we can distinguish probably two modifications in cordierite, regardless of the Δ value. They are called *low and high cordierites*. High cordierite has much lower refractive indices than low with the same Δ value and composition. The structural change from low to high cordierite also appears to take place gradually through intermediate states.

Initially the Mont Bity mineral is low cordierite, whereas the Asama and Kasyo-to minerals are high cordierite formed at fairly high temperatures under volcanic conditions, and hence they do not show any decrease in β index on heating. Natural cordierite emits water on heating. The relation of this fact to the high-low transition was discussed in another paper (Miyashiro and others, 1955, p. 200-201).

Laramie Range cordierite no. 2.—This cordierite occurs in a metasomatic deposit in meta-norite in the Laramie Range, Albany County, Wyoming (Newhouse and Hagner, 1949). As shown in table 2, it has a very low Fe^{+2}/Mg ratio. It is suitable for heating experiments for long periods at high temperatures, for the oxidation of iron has probably little effect on the properties of this mineral.

The thermal behavior of this mineral is similar to that of the Mont Bity (table 6). Initially it is in a metastable state with Δ = 0.27. On heating at 1000° and 1100°C, it changes to the stable state with Δ = 0.29. On heating at and above 1200°C, the mineral resumes subdistortional-cordierite states with gradually smaller Δ values, which are stable at those temperatures. Probably it becomes indialite eventually at a temperature above 1420°C. (Recently

TABLE 6
Heating Experiment of the Laramie Range Cordierite No. 2

Temperature (°C)	Time of heating (in minutes)	Δ	2V about X	β_D
Room temp.		0.27	77°	1.541
1000	10	0.29	77°	1.530
1000	60	0.29		1.531
1100	60	0.29		
1200	60	0.27		1.530
1400	10	0.24		
1400	60	0.24		
1400	240	0.24		
1420	240	0.21		

Sugiura and Kuroda (1955) made a heating experiment of this specimen. They declared that this cordierite changed to indialite on heating at 1400°C. However, judging from the X-ray diffractogram shown in figure 2 of their paper, it appears to me that the material they obtained by the heating is not indialite but actually a mixture of subdistortional cordierites with variable Δ values.)

The β index decreases on heating. Before heating, $\Delta = 0.27$ and $\beta = 1.541$. After heating at 1200°C, $\Delta = 0.27$ and $\beta = 1.530$. These two samples, with the same Δ value, have different β index. Therefore, it is considered that the initial material is low cordierite, and the heated sample is high cordierite.

Table 6 shows that the mineral reaches the stable state by heating only for 10 minutes. Further heating for a longer period does not cause any more noticeable change in Δ value.

The Laramie Range cordierite no. 1 with a higher Fe^{+2}/Mg ratio (FeO = 7.06%) shows a decrease of about 0.005 in the refractive indices on heating at 1000°C for 10 minutes (Miyashiro and others, 1955, p. 198).

Subdistortional Cordierites with $\Delta < 0.20$

Haddam cordierite.—This cordierite occurs in pegmatite cutting through biotite gneiss at Haddam, Connecticut (Heinrich, 1950, p. 177-178). The chemical composition is shown in table 2.

Table 7 and figures 4 and 5 show the results of the heating experiment. They may be interpreted as follows: Initially the cordierite is in a metastable state with $\Delta = 0.12$. On heating at 700°-1200°C, it changes into states with larger Δ values, which are more stable at those temperatures. The Δ value of the heated samples is largest after heating at 1200°C. However, the state of perdistortional cordierite has not been reached in these heat treatments. Eventually on heating at 1360°C, it becomes indialite. On heating at 1380°C, it decomposes into liquid and mullite.

On heating, the β index decreases gradually probably owing to the two kinds of structural change: one from low cordierite towards high and the other of the changing Δ type.

TABLE 7
Heating Experiment of the Haddam Cordierite

Temperature (°C)	After heating for 10 minutes, 2V about X: Δ		β_D	After heating for 60 minutes Δ
Room temp.	0.12	42°	1.560	0.12
600	0.12		1.560	0.12
700	0.12		1.560	0.145
800	0.15		1.547	
900	0.18		1.542	0.15
1000	0.18	66°	1.542	0.19
1100	0.19		1.542	
1200	0.20		1.541	0.20
1250	0.19			0.175
1300	0.18			0.10
1330	0.10			Very small
1360	0.00			0.00
1380	Decomposes			Decomposes

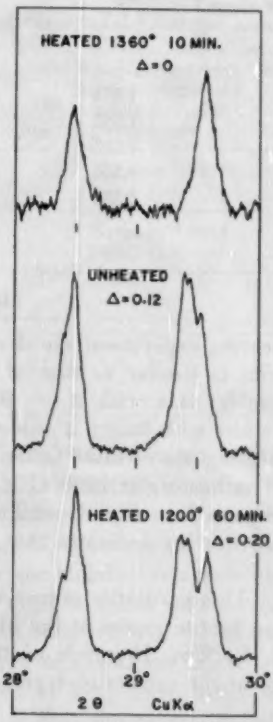

Fig. 5. X-ray diffractograms of unheated and heated samples of the Haddam cordierite.

The state of perdistortional cordierite is much more difficult to reach in the heat treatment of subdistortional cordierite with $\Delta<0.20$ than in that of subdistortional cordierite with $\Delta>0.20$.

The amount of decrease in β index on heating is much larger in subdistortional cordierites with $\Delta<0.20$ than in subdistortional cordierites with $\Delta>0.20$. For example, it is 0.019 in the Haddam cordierite, whereas it is of the order of 0.005-0.011 in subdistortional cordierites with $\Delta>0.20$, as shown before.

Sugama cordierite.—This cordierite occurs in a granitic pegmatite cutting through gneissic hornblende-biotite granodiorite and metamorphic rocks at Sugama near Isikawa, about 200 kilometers north-northeast of Tokyo (Sambonsugi, 1956). Cordierite samples from various parts of the pegmatite mass do not show any noticeable difference in the Δ value. The cordierite has a very high Fe^{+2}/Mg ratio, as shown in table 2.

TABLE 8

Heating Experiment of the Sugama Cordierite

Temperature (°C)	After heating for 10 minutes Δ	2V about X	β_D	After heating for 40 minutes Δ	After heating for 60 minutes Δ
Room temp.	0.125	55°	1.574	0.125	0.125
500	0.125			0.125	0.125
700	0.135		1.574	0.135	0.135
800	0.22		1.555		
900	0.22		1.555	0.22	0.26
1000	0.22	67°	1.555		
1100	0.22		1.555	0.20	
1150	0.15			0.17	
1175				0.05	
1200	0.00			0.00	
1250	Decomposes			Decomposes	

The results of the heating experiment are shown in table 8. The thermal behavior of this cordierite is similar to that of the Haddam. Initially the cordierite is in a metastable state with $\Delta = 0.125$. On heating at 700°-1000°C, it changes into states with larger Δ values, which are more stable at those temperatures. On heating above 1100°C, the Δ value becomes gradually smaller again. Eventually on heating at 1200°C, it becomes indialite. On heating at 1250°C, it decomposes into liquid, mullite, etc. The remarkable decrease in the β index on heating indicates that the initial material is low cordierite.

Micanite cordierite.—This cordierite occurs in a granitic pegmatite cutting through Precambrian biotite gneiss at the Micanite district in Colorado (Heinrich, 1950). Prof. E. Wm. Heinrich of the University of Michigan kindly gave me specimens of the cordierite, together with a cordierite-bearing country rock.

The pegmatite cordierite shows $\Delta = 0.13$. On heating at 1000°C for 20 minutes, the Δ value becomes 0.22. The β index is 1.568 before heating and 1.549 after heating.

It is of some interest to note that the cordierite in the country rock (sillimanite-cordierite-biotite gneiss) shows $\Delta = 0.26$.

STABILITY RELATION OF CORDIERITE AND INDIALITE

According to the heating experiments of the preceding chapter, perdistortional cordierite ($\Delta = 0.29$-0.31) represents the state stable at low temperatures, and subdistortional cordierites are stable at higher temperatures. At still higher temperatures, indialite ($\Delta = 0$) is stable. Eventually at a very high temperature, indialite decomposes into liquid and crystals (mullite and probably in some cases tridymite). The structural changes take place by heating only for 10 to 20 minutes in many cases.

The temperatures of structural changes and decomposition become lower with increasing Fe^{+2}/Mg ratio as shown in table 9. Judging from the studies of Rankin and Merwin (1918) and of Schairer and Yagi (1952),the decomposition temperatures of pure Mg- and Fe-indialites are 1470° and 1210°C respectively. Thus, we obtain the equilibrium diagram of the system $Mg_2Al_4Si_5O_{18}$—$Fe_2Al_4Si_5O_{18}$ as shown in figure 6.

TABLE 9
Temperatures of Structural Changes and Decomposition

	$\frac{(Fe^{+2}+Mn)\times 100}{Mg+Fe^{+2}+Mn}$	Boundary between perdistortional and subdistortional cordierite	Boundary between subdistortional cordierite and indialite	Decomposing temperature
Laramie Range cordierite no. 2	11.1	Between 1100° and 1200°C	Above 1420°C	?
Mont Bity cordierite	22.2	Between 1000° and 1100°C	Slightly above 1415°C	?
Haddam cordierite	33.6	?	1350°C	1370°C
Sugama cordierite	70.3	?	1180°C	Between 1200° and 1250°C

This diagram gives some informations on conditions of formation of cordierites and indialite. However, it must be emphasized that there are reasons to believe that various structural modifications of cordierite and indialite are formed metastably in many cases in artificial processes as well as in nature. Thus, not only the stability relations but also the mechanism and velocity of crystallization must be taken into consideration for understanding the modes of formation and occurrence of the minerals of the cordierite group.

The manner of change of distortion index with temperature, as shown in figure 4, resembles that of long-range order with temperature in alloys. This fact supports the view that the cordierite-indialite transformation is of the order-disorder type, and the distortion index is a good measure of the degree of order.

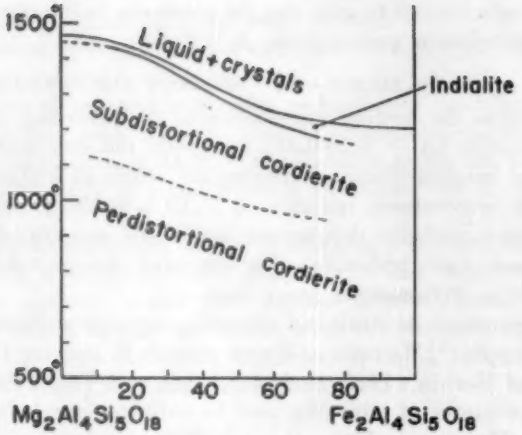

Fig. 6. The equilibrium diagram of the system $Mg_2Al_4Si_5O_{18}$—$Fe_2Al_4Si_5O_{18}$. (The distinction between high and low modifications is not shown in this diagram. "Indialite" and "subdistortional cordierite" of this diagram represent the high modifications, while "perdistortional cordierite" represents both the high and low modifications.

HIGH AND LOW CORDIERITES

In a previous section I pointed out evidence to indicate that the materials from volcanic rocks of Asama and Kasyo-to are high cordierite, while the materials from metamorphic rocks, pegmatites, and quartz veins are low cordierite. (In this paper the term metamorphic rocks does not connote the inclusions in igneous rocks.) Corroborative evidences on this problem will be discussed below.

If we compare the Haddam and Sugama cordierites heated around 900°-1200°C with low cordierites with $\Delta = 0.26$ or so of similar compositions, the former minerals have smaller Δ values (0.18-0.22) and much lower refractive indices than the latter ones whose β index is shown in figure 3. This is contrary to the tendency that the smaller the distortion index of cordierites from metamorphic rocks, pegmatites and quartz veins, the higher the refractive indices. This fact cannot be explained until we recognize that the heated samples belong to the group of high cordierite which has much lower refractive indices than cordierite from metamorphic rocks, pegmatites, and quartz veins (i.e. low cordierite), when they have the same chemical composition.

Probably the transition from low to high cordierite also is of order-disorder type and takes place gradually through intermediate states. The reverse change has not been noticed in my experiments. It is interesting that this transition has only a very little effect on the optical angle, as shown in figure 2.

The refractive indices of low cordierite increase with decreasing Δ. Therefore, indialite ($\Delta = 0$) should have higher refractive indices than low cordierite. However, we know that indialites synthesized in dry melt experiments (i.e. the α-forms of $Mg_2Al_4Si_5O_{18}$ and $Fe_2Al_4Si_5O_{18}$) have much lower refractive indices than low cordierites of the same compositions. Therefore,

we must suppose another form of indialite ($\Delta = 0$) with much higher refractive indices than the above-mentioned form of indialite, as the limiting case of decreasing Δ value in low cordierite. Thus, the form of indialite from dry melt (the α-form) may be called *high indialite*, whereas another form of indialite with much higher refractive indices may be called *low indialite*. Through intermediate states, high indialite is probably continuous to high perdistortional cordierite, whereas low indialite may be supposed to be continuous to low perdistortional cordierite.

Yoder (1952) synthesized what appears to be high subdistortional cordierite ($2V = 30°$) by the hydrothermal method at temperatures above 830°C, while he obtained similar crystals ($2V < 45°$) with much higher refractive indices by the same method at temperatures below 830°C. The latter is probably a low subdistortional cordierite with a small Δ value. (This material was called the β-form by Karkhanavala and Hummel (1953).)

The high-low transition point of cordierites is probably lower than 830°C, at least in most cases. Therefore, we come to the conclusion that low perdistortional cordierite, high perdistortional cordierite, high subdistortional cordierite, and high indialite are probably stable at successively higher temperatures. Low indialite and low subdistortional cordierite are probably unstable in all cases.

The relation between the refractive indices and chemical composition in high cordierite is discussed in other papers (Miyashiro and others, 1955, fig. 2; Iiyama, 1956).

FORMATION OF LOW SUBDISTORTIONAL CORDIERITE

All the examined cordierites from metamorphic rocks, pegmatites, and quartz veins belong to the class of low subdistortional cordierite, which is unstable as noted above. Probably they were formed as a metastable phase at temperatures so low that the change from the metastable to the stable state could not take place during and after their formation. Low subdistortional cordierites with Δ values smaller than 0.20 occur characteristically in pegmatites, and have Fe^{+2}/Mg ratios larger than 3/7, so far as I am aware. The rapid growth of big crystals and higher Fe^{+2}/Mg ratios in pegmatites may favor the formation of metastable cordierites with such small Δ values.

The lowest limit of temperature at which natural subdistortional cordierite changes into perdistortional on heating should represent the highest possible temperature of its formation. Tables 5, 7, and 8 show that such a change takes place at a temperature as low as 700°C within 1 hour in those cordierites, whereas the Haddam cordierite, heated at 600°C for 7 hours, shows little appreciable change in Δ. These data suggest that the metastable cordierites were formed at temperatures lower than 700°C.

There is, however, a serious defect in this reasoning. The pressures of water vapor and oxygen, under which natural cordierites are formed, differ from those of water vapor and oxygen in the laboratory. Therefore, the manner of emission of water and oxidation on heating in the laboratory must differ from the corresponding phenomena in nature. This difference must have some effect on the temperature and ease of structural changes. Since

the magnitudes of these effects are not known, it is impossible to determine the temperature of formation of natural cordierite by simple heating.

The conditions and mechanism of formation of metastable cordierite will be discussed in a later paper.

CONCLUSION

I am presenting below a revised structural classification of the members of the cordierite group of minerals on the basis of the investigation described in this paper.

In order to understand the variations in the structural states of these minerals, the following four limiting forms are to be set forth:

(1) High indialite (identical with the α-form synthesized by Rankin and Merwin, 1918). It is hexagonal ($\Delta = 0$) and stable at very high temperatures. It occurs in fused sediments of Bokaro coalfield in India.

(2) Low indialite (supposed). It is hexagonal ($\Delta = 0$) and unstable at all temperatures. It has been found neither in nature nor in artificial products.

(3) High perdistortional cordierite. It is pseudohexagonal ($\Delta = 0.29$-0.31) and stable at medium temperatures. It was found in xenoliths in andesite from Asama, and also in blast-furnace linings.

(4) Low perdistortional cordierite (supposed). It is pseudohexagonal ($\Delta = 0.29$-0.31) and stable at low temperatures. It has been found neither in nature nor in artificial products, though I expect that it will be found in some metamorphic rocks.

There exist many materials in structurally intermediate states between indialites and perdistortional cordierites. Thus, most of the natural as well as artificial "cordierites" belong to either of the following two classes:

(5) High subdistortional cordierite (intermediate between high indialite and high perdistortional cordierite). It is pseudohexagonal ($0.29>\Delta>0$) and stable at high temperatures. It was found in andesite from Kasyo-to, and also in fused sediments from Bokaro coalfield. Most cordierites in industrial materials may belong to this class.

(6) Low subdistortional cordierite (intermediate between low indialite and low perdistortional cordierite). It is pseudohexagonal ($0.29>\Delta>0$) and unstable at all temperatures. Most cordierites, if not all, in ordinary metamorphic rocks, pegmatites, and quartz veins belong to this class.

The nature of the μ-form of Rankin and Merwin (1918) is not clear. What was called the β-form by Karkhanavala and Hummel (1953) is probably a low subdistortional cordierite with a small Δ value. In this classification I have ignored the existence of intermediate states between high and low modifications in cordierite as well as in indialite, because otherwise the classification would become too complicated.

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MASS WASTING AND THE DEFORMATION OF TREES

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ABSTRACT. The concept of slow mass wasting, or soil creep, is allegedly supported by the displacement of surface objects. Curvature of tree trunks is a widely used manifestation of the process, although in the search for evidence of soil creep the more logical deforming agents are disregarded and the physiology of tree growth has been ignored.

This paper presents the mechanical and physiological causes for tree deformation and discusses the type of curvature which would develop under concomitant tree growth and soil creep. It is concluded that the proper tree curvature is rarely, if ever, observed, and that alternative explanations for the development of the curvature which does exist are probably more valid. With doubt cast upon one of the principal evidences of soil creep, it seems pertinent to subject the other criteria to a more thorough examination and perhaps re-evaluate or redefine soil creep as a geomorphic process.

INTRODUCTION

Most geomorphic processes operate so slowly that the mechanics must be inductively postulated, and the evidence proposed to support these processes is, therefore, frequently drawn from non-geologic fields. Such being the case, the evidence must be critically tested by principles of applicable scientific disciplines, and thus "the practicing geomorphologist will be more effective if well versed in soil science and acquainted with principles of hydraulics, meteorology, oceanography, and botany" (Sharp, 1950). One widely accepted concept that has not passed the necessary test and which becomes less reasonable and possibly invalid by critical examination is the presumed correlation between tree curvature and soil creep.

Soil creep,[1] having received almost universal acceptance as an important degrading agent, is included in all geomorphology texts. Phenomena cited to support soil creep include the bending of trees, dislocation of surficial objects, and such subsurface features as warped beds and lines of stones (fig. 1). The bending of trees is in most cases emphasized as indicative of downslope movement or slow flow of soils. One author states: "The movement is very slow, but can be measured as it influences the growth of trees. The root system of these is anchored in the deeper horizons which move more slowly, or in the living rock itself, whilst the quicker moving superficial layer presses the upper parts of the root system downslope. This rotation couple makes the trees oblique at the base, while the trunks as a whole tend to grow vertically upwards. Thus they show curvature, convex on the downslope side" (Penck, 1953, p. 109). Another states, "Downhill tilting of trees and a compensating upslope curvature of their trunks is unusually well developed on hillsides of relatively loose material . . . " (Sharpe, 1938, p. 24). A third says, "The trunks of trees are sometimes tilted by the downslope displacement of the soil about their roots, and they may even develop a curved shape, bending convexly down the slope

[1] Although the term creep was used by Davison in 1889, the most definitive and widely quoted study of mass wasting was made by Sharpe (1938). In this study soil creep was defined by Sharpe (p. 21) as the "slow downslope movement of superficial soil or rock debris, usually imperceptible except to observations of long duration"; he concluded that the slow downhill creeping process is continuous although partially due to an infinite succession of very minute movements. Unfortunately, the term soil creep has come to encompass a wider variety of gravitative processes than was seemingly intended, until now it is frequently used as a catch-all term or a convenient pedagogical tool.

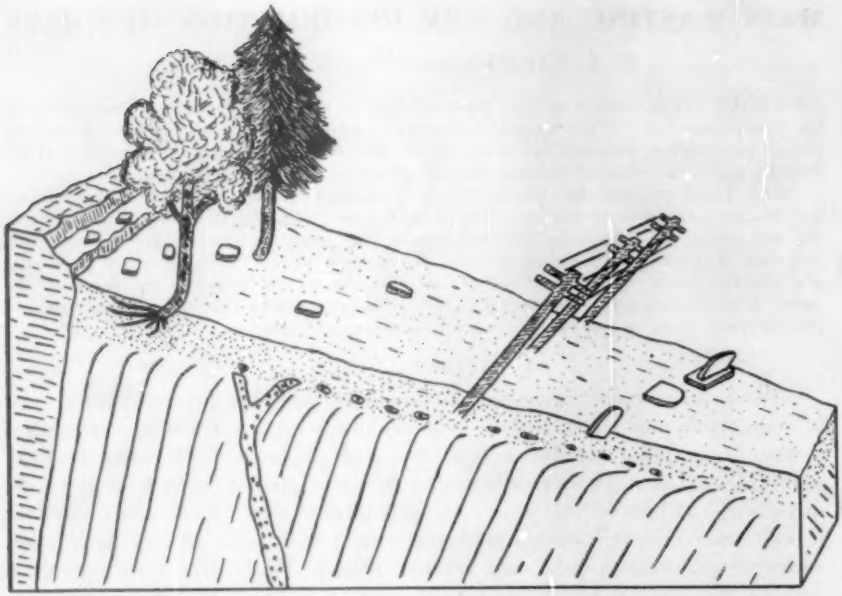

Fig. 1. The conventional evidence of soil creep, showing displacement of surface and subsurface objects.

in their efforts to grow into the normal vertical position" (Monnett and Brown, 1950, p. 60). Similar statements can be found in other geological texts, although in botanical writings soil creep is disregarded as a contributing factor to tree curvature during growth. It would appear that the willing acceptance by many geologists of distorted trees as an evidence of soil creep, without examination of the physiology of the growth, has been, in effect, search *for* rather than examination *of* the evidence. It is pertinent and timely, therefore, to examine and evaluate this correlation which apparently is so firmly imbedded in the minds of many.

HYPOTHETICAL TREE CURVATURE

Without exception, illustrations of trees allegedly bent by creeping soil portray one sharp curve above which the bole is vertical. The curve supposedly results from pressures set up by faster-moving surficial soils against the roots of the tree which are anchored in deeper slower-moving soils or in the bedrock. The downslope tilt of the tree resulting from these stresses is corrected by geotropic response, the renewed vertical growth producing one relatively angular curve. But if soil creep is a slow but continuous movement as defined, and downslope bending of trees the result, there should be no straight segment of the tree stem; rather, the trunk of the tree should assume under these

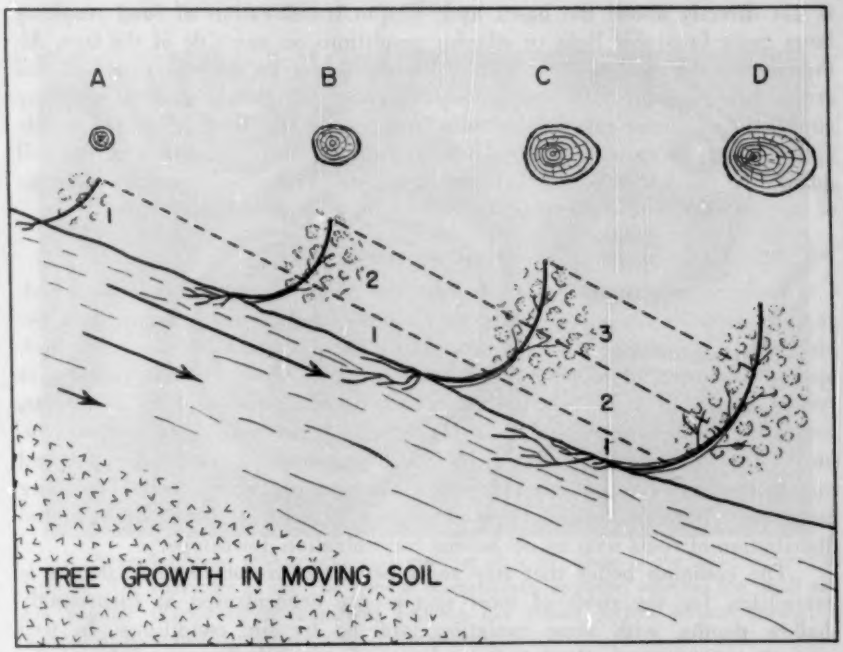

Fig. 2. The hypothetical sequence of tree growth in slow, continuously moving soil. Of significance is the single sweeping curve from terminal bud to base and the asymmetrical growth rings.

conditions a long continuous curve with only the terminal buds attempting to maintain a vertical attitude. Figure 2 illustrates the hypothetical sequence of tree growth in a continuously slow-moving soil. The curvature produced by initial movement (A-1, fig. 2) must persist throughout the life of the tree, since growth is restricted to the peripheral layer, and only the terminal buds respond geotropically. Significantly, the deformed tree remains curved, for it is only a board, and the initial curvature produced in stage A-1 must be constant in degree and relative position throughout the life of the tree. This is demonstrated (fig. 2) in the succeeding downslope positions (B, C, D) of the tree. The segments of the stem below the broken line extending from the crown in A-1 to the base of the trunk in stage D are identically curved, differing only in that they are increasingly canted downslope. Later bending, represented by B-2, C-3, and in D, are upward extensions of the already deformed bole, the degree of curvature reflecting the rapidity of downslope movement of the soils and tree. Carried to the extreme, the first and lowest curve (D, fig. 2) would be recumbent.

Cross sections of these curved trees (fig. 2) should have asymmetrical growth rings which are increasingly wider on the downslope side of the tree. Botanists generally credit this asymmetry of growth to (1) gravitational concentration of growth hormones, the added wood or compression tissue on the downslope side of the trunk strengthening the tree, the bulk of whose weight

is not directly above the base; and, (2) to translocation of food resulting from more favorable light or edaphic conditions on one side of the tree. Although this feature in itself would, therefore, not be definite proof of soil creep, trees deformed by continuously moving soil should show a sweeping curve and asymmetrical cross sections from tip to base. Vertical stem segments and symmetrical growth rings would indicate a static condition of the soil during the periods in which this growth was occurring. Such lengthy intervals of soil stability would preclude, by definition, a condition of creeping soils.

ROOT PATTERNS

Because tree curvature is allegedly the result of roots anchored in bedrock or more slowly moving soils, they are depicted as strung out upslope behind the moving tree (fig. 2). Accepting the corollary of tree curvature, upslope extension of roots, and soil creep, a tree with an initially radial root system would, in a creeping soil, encroach upon a portion of its own roots, contorting and concentrating them beneath the trunk until as movement continued they became unraveled and drawn out upslope. Logically, old and mature trees growing on active slopes should have few, if any, roots extending downslope. However, examination of trees on even steep slopes shows radial distribution of roots with an occasional concentration downslope.[2]

The common belief that tree roots penetrate to considerable depths is unrealistic, for the roots of most species are concentrated at surprisingly shallow depths, with some variation imposed by site conditions. In well stocked mature stands the intertwined roots form a shallow tangled mat developed so compactly that movement of individual trees is impeded, if not precluded. Where minor slumping or other rapid movement disturbs the forest, the mat of roots and trees above moves as a unit, similar to a toboggan. If the movement is moderate both in distance and velocity the trees may remain vertical or be unevenly tilted; but should the mass movement be violent and extensive, the trees are felled and the forest destroyed.

SUBSTANTIATED CAUSES FOR TREE CURVATURE

Although deformed stems are characteristic of all woodlands, there is usually no common denominator of degree, direction, or position of curvature (pl. 1-A), and the hypothetically curved trees (pl. 1-B) are significantly absent. At each site more careful observation suggests that factors other than

[2] Illustrations in a paper by Helmers et al., (1955), display root systems growing on considerable slopes in Southern California. In all cases a radial pattern is developed, and exposure of the roots of chaparral yucca showed that "an equal number of roots grew in all directions from the plant, but those growing downhill extended 11 feet while those growing uphill extended less than 6 feet" (p. 675). It should be noted that the authors refer to roots growing uphill and no reference is made to roots being strung out upslope by soil creep. No uphill dragging of the roots is exhibited in an illustration of Eastwood manzanita, although they penetrate more than 9 feet into the bedrock. In fact, this species is recommended by the authors as a plant useful in preventing mass movements on oversteepened slopes. This seemingly contradicts the statement that trees anchored in bedrock move downslope, leaving their roots trailing behind. It appears probable that if roots anchored in the bedrock help stabilize the slope, then soil creep would be minimized or non-existent at such localities; but if the roots are not anchored, soil creep would produce no tilt of the trees and, therefore, no opportunity for the supposed curvature to develop.

PLATE 1

A. Short leaf pine (*Pinus echinata*) growing on a 20° slope in the mid-Piedmont of Georgia. The curvature of the two trees is diametrically opposite, whereas the small yellow popular (*Liriodendron tulipifera*) between has almost perfect vertical growth. A number of deformed stems are visible in the background.

B. Large loblolly pine (*Pinus taeda*) in background and persimmon (*Diospyros virginiana*) in left foreground are growing vertically on a 30° slope composed of loose fill. The trees are 60-80 years old. Location is 2.5 miles northeast of Athens, Clarke County, Georgia.

soil creep are undoubtedly more critical in the deformation. More realistic causes can be broadly classified as mechanical and physiological. The mechanical causes include the effects of unusual weather conditions, i.e., high winds, ice, snow, damage during logging, and obstructions to growth from rock outcrops. The physiological factors result from gravity working upon trees whose supporting roots have been weakened in any way, i.e., pathogens, fires, etc., to destruction of the terminal buds in any manner, to growth responses to light or food conditions, and to vegetative regeneration. It is evident that rarely, if ever, is the aerial portion of a tree perfectly balanced, enhancing the probability that it will topple under certain circumstances. In fact, the ultimate fate of the majority of trees is to be felled by this imbalance in conjunction with one of the weakening agents.

Omnipresent pits and mounds on forest floors, produced by displaced roots as the trees were tilted or felled, testify to the past and continuing destiny of all trees. The disturbance of soils by the upheaval of roots was noted by Shaler (1891), Holmes (1893), Van Hise (1904); Lutz and Griswold (1939) demonstrated clearly the profound changes in the soil profile accompanying this process. A detailed study of mounds and pits in the Harvard forest by Stephens (1956) reveals that "the area has been subjected to four periods of major uprootings . . . . The fourth and oldest was estimated to have occurred between 1400 and 1500 from evidence existing solely on the ground" (p. 115). It is notable that in an area of active frost heaving, a condition said to be conducive to soil creep, the gross anatomy of mounds and pits has been retained for such lengthy periods. Significantly, soil creep is disregarded by Stephens in the discussion of contributing causes for the throwing of trees, and he concludes that "uprooting assumes the proportions of a general process of the forest; and . . . can be considered in the same way as other generally recognized processes, such as growth, reproduction, and podzolization" (p. 116). With complete felling of trees playing this widespread role in forests, the same factors would doubtless produce even greater numbers of partially felled or tilted trees.

ANALYSIS OF SPECIFIC SITES

Examination of a specific illustration of tree curvature used to support soil creep (Sharpe, 1938, p. 27) substantiates the belief that the correlation is tenuous and that alternate causes for tree deformation are more significant. The sketch (fig. 3) is of a photograph taken in Lassen Volcanic Park, California. Here the conifers are growing in relatively loose volcanic soils, with each tree possessing a single curve occurring at a uniform height, above which the trunks are vertical. Considering the evidence, a number of causes other than soil creep are seemingly more valid: (1) While they were still saplings a sudden slump of surficial soils and root mat tipped the trees downslope, after which they righted themselves geotropically, producing a straight bole above a single bend. Similar age and height within the stand at the time of slump is suggested by the uniform position and direction of curvature. Although the circumstances are unknown, perhaps oversteepening of the slope during road construction explains the sudden slump of the inherently unstable vol-

Fig. 3. Sketch of conifers in Lassen Volcanic Park, California. Taken from photograph used by Sharpe (1938) to illustrate the effects of soil creep upon trees.

canic soils. (2) Clearing the right of way for the road, if not directly responsible for the slump, may have changed the light conditions and provided the impetus for phototropic responses and non-vertical growth of saplings growing in the shade of an overstory. (3) Also possible and highly probable at this site are unusual weather conditions, including severe ice and snow loads, or gale winds. One of these mechanical agents may have deformed the trees while still saplings.

Not only has reconnaissance of the forested slopes of the Georgia Piedmont failed to reveal tree curvature that can be definitely attributed to soil creep but the feature is also absent in specific sites where ideal conditions are apparently fulfilled (Parizek and Woodruff, 1956). For example, the 30° slope (pl. 1-B) developed by the artificial fill for a reservoir dam is composed of loose material that has been in situ for more than 80 years. The long straight boles of the mature forest attest to the stability of the soil at this locality, or to the ineffectiveness of soil creep in tree deformation.

SUMMARY

Trees curved concave downslope have long been held to indicate soil creep; but examination in the light of botanical knowledge contradicts such a correlation. Concomitant tree growth and continuously slow-creeping soil should hypothetically develop trunks with a long sweeping curve from ground to terminal bud rather than the elbow-type curve and vertical stem segments believed to manifest soil creep. The widespread occurrence of trees with the latter type of deformation can be more validly related to mechanical and physiological causes.

However, granting that a correlation may exist between deformed trees and soil creep, detailed study suggests: (1) soil creep is much less important than believed because uniformity of curvature in stands is the exception and not the rule; or, (2) if soil creep is widespread, then there must be a counteracting process or unknown soil conditions which effectively prevent the development of the properly related tree curvature.

Because tree curvature as supporting evidence of soil creep is, therefore, tenuous or at least over-emphasized, the relevancy of other displaced objects of soil creep should be tested by principles of applicable fields. The results of further investigation may require re-evaluation or redefinition of the concept of soil creep.

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ESSAY REVIEW

Cenozoic Geology of the Colorado Plateau; by CHARLES B. HUNT. p. 99, 62 figs. Professional Paper 279, U. S. Geological Survey, Washington, D. C., 1956 (75 cents paper cover).—The Colorado Plateau offers exceptional exposures of bedrock, because of deep dissection, generally simple structure, and semiarid climate. From early Paleozoic until early Cenozoic time a major part of that wide region was below or only slightly above sea level; accordingly for a long succession of geologic periods an outline of the history is recorded in sedimentary deposits. But for the region as a plateau the history is in large part obscure, because repeated uplift prevented extensive sedimentation and brought about at least partial destruction of deposits that were formed locally. Attempts to piece together the record since Eocene time must use fragmentary evidence found within the plateau itself, supplemented by records preserved in adjoining regions. Hypothesis is required to make a connected story, and only a geologist with wide acquaintance in the plateau country is competent to fit the known facts into credible speculations. Charles B. Hunt is well qualified for the task from his long acquaintance through field study in the Mount Taylor area, the Henry Mountains, and other parts of the Colorado Plateau. His first-hand information is supplemented by knowledge of a literature that is surprisingly extensive; the bibliography published with the paper lists about 400 titles.

Hunt's lengthy essay starts properly by giving a good geographic setting, with maps and diagrams supplementing the text. Even the average geologist who has some acquaintance with the Colorado Plateau will welcome the sketch maps showing in proper relation the numerous topographic and geologic features that are important in the textual discussion. The pre-Cenozoic geology —Precambrian, Paleozoic, and Mesozoic—is summarized in about four pages. Paleocene and Eocene formations, which are extensive and give definite information on paleogeography, are discussed in about nine pages. Discussion of the younger Cenozoic sedimentary and igneous rocks occupies about 25 pages, because the number of units is large and uncertainties in dating and correlation require explanation. No doubt the uncertainties will surprise many readers. In the correlation chart, Fig. 12, 18 of 21 sedimentary units younger than Eocene are placed with question marks. Intrusive igneous bodies, as well as many units of volcanic rock, also are of very dubious age, with estimates resting largely on geomorphic relationships. Even the sequence of many events, as well as the dating, has to be stated with reservations. Recent discovery of the Bidahochi formation, dated by vertebrate fossils as Pliocene, raises hopes that other helpful additions to stratigraphic knowledge are still to come, at least through the finding of diagnostic fossils in deposits now known but undated.

Structural features developed in late Cretaceous and early Cenozoic time are dated with fair certainty by sedimentary evidence. In this category are the monoclinal folds such as the East Kaibab and Echo Cliffs, and broad anticlinal arches like that at Circle Cliffs. Later crustal movements include displacements on faults and broad warping, but exact dating of events is still a major goal. Presumably a large part of the plateau uplift occurred late in Cenozoic time, but in what epoch or epochs? Hunt notes that large-scale

movement on the Grand Wash fault, probably connected with important rise of the plateau, occurred during deposition of coarse deposits tentatively correlated with the Muddy Creek formation. On this basis the uplift in question is assumed to have occurred "in late Miocene and early Pliocene". Unfortunately no part of the Muddy Creek formation is firmly dated, and the critical deposits in the Grand Wash trough are not assuredly within that formation. Thus uncertainties in the chronology are compounded.

The history of geomorphic features in the plateau has long been a subject of controversial speculation. Powell, and later Dutton, supposed that the Colorado River maintained its course across rising blocks bounded by the East Kaibab monocline and large faults west of it. According to this concept the course of the river in the Grand Canyon would be classed as antecedent. Later, when it became known that early Cenozoic deposits lie with angular unconformity on the bevelled monoclines, W. M. Davis suggested that the course of the river, established on the younger formations, became superposed on the deformed older rocks during uplift of the plateau. Incised meander patterns of the Colorado and its tributaries were cited as evidence of the superposition. Hunt now suggests a combination of antecedence and superposition. He points out that the Grand Canyon section of the plateau is structurally high; and he proposes that in Miocene and later time the flow of the Colorado River was obstructed by differential uplift in this section, involving renewed growth of the Kaibab upwarp. Resultant ponding farther upstream must have caused widespread deposition; and, when the river eventually deepened its valley in the area of uplift, its upstream course was cut down through the fill and became superposed on the older bedrock. This general concept is not new. Localized uplift athwart a stream course must inevitably cause ponding upstream, thus creating conditions favorable for antecedence and superposition in contiguous segments of the valley. An actual example of such a history, described by W. C. Putnam, is cited by Hunt. But application of the concept to the Colorado River history is new, and Hunt urges several arguments in its favor. Obstruction of the river while the Muddy Creek deposits accumulated in interior basins will explain the absence of record made by a through-flowing stream west of the plateau in Muddy Creek time; the assumed history explains the classic examples of incised meanders upstream from the Grand Canyon; obstruction of the drainage by renewed uplift in late Cenozoic time offers a plausible explanation of the Bidahochi lake and stream deposits in northeastern Arizona.

For the full process involving antecedence and superposition in adjacent parts of a stream course, Hunt proposes the new term *anteposition*. In the concept itself I see considerable merit. The wide plateau was raised in late Cenozoic time, unequally in its different parts and probably in a succession of pulses. Large movements occurred on the Hurricane and other faults, and it is wholly credible that the rise of blocks athwart the Grand Canyon repeatedly slowed or actually obstructed the flow of the Colorado River. If this occurred on the scale suggested by Hunt, we may hope to find corroboration in remnants of Tertiary deposits older than the Bidahochi. The hypothesis bears on several regional problems and therefore merits thorough testing. But my en-

thusiasm for the concept does not extend to the name proposed for it. The suggested mechanism involves two distinct processes, antecedence and superposition, operating in logical unison. These two terms are good words in their own right, and have geologic implications established by long usage. Joining fragments of the two words to make another that has no obvious meaning is bad etymological practise. And as the hybrid word can be understood only in terms of the two for which it is meant to do double duty, it has doubtful value at best and its use may serve to confuse rather than clarify. This is illustrated by Hunt's references (p. 67, 70) to drainage that is "apparently superimposed or anteposed (or both)". If by definition "anteposition" includes superposition as an essential ingredient, how can the two processes be recognized as distinctive and independent? Cluttering of our geologic vocabulary with superfluous special terms should be resisted.

Hunt's paper closes with a frankly hypothetical outline of Cenozoic history of the plateau, embellished with ten sketch maps in relief that illustrate stages in development from late Cretaceous to Recent time. Preparation of such maps requires considerable courage, for they involve definite commitments on controversial matters; and as W. M. Davis once commented, "An author may suffer more from his diagrams than from his ideas". But without question Hunt's maps help greatly in making his main concepts clear, and they enable readers to keep in mind relationships among the numerous major features in the vast plateau and its immediate surroundings. The maps bring into sharp focus the passing of topographic dominance, after early Cenozoic time, from the Basin and Range area to the Colorado Plateau. Two hypothetical points will stir particular interest: representation (Fig. 60) that the Kaibab upwarp was much extended southward in mid-Miocene time; and (Fig. 62) that the lower 40-mile stretch of the Grand Canyon has been excavated (through stream diversion) since the middle of the Pliocene epoch. Both hypotheses are challenging and call for further field study.

In some respects this professional paper does not measure up to the standard of careful editing expected of works published by the Geological Survey. The author uses *superimposed* and the shortened form *superposed* interchangeably and often; on pages 65-71 the two forms are curiously alternated. Good arguments may be urged for preferring one form or the other, but readers expect consistency in any one publication. The correlation chart, Fig. 12, is misleading in its listing the Muddy Creek formation and Hualpai limestone among sedimentary units of "northeast Arizona"—actually neither formation, as now known, is found farther east than extreme north*west* Arizona. The author makes effective use of numerous diagrams copied from other publications; but unfortunately some are reproduced with errors. A section published by Darton, extending from Holbrook southward through Pinedale, is almost exactly 50 miles long, but the reproduction in Hunt's Fig. 11 states the length as "about 90 miles". In the map of Fig. 22, the Iceberg fault is shown erroneously as cutting Tertiary beds; *Grapevine Wash* appears both north and south of the Colorado River (suggesting that these stream courses are mutually transecting?); and the *Explanation* misstates the identity of Paleozoic strata. Readers familiar with the Paleozoic section of the region

will be astonished on seeing in Fig. 23, A the Cambrian Bright Angel (Pioche) shale represented as a coarse fanglomerate and indistinguishable from a Tertiary unit in the same section. For many readers this error destroys the value of the diagram. Because details of a drawing register at a glance, incorrect reproduction can be more damaging than misquotation of text.

In an overall appraisal Hunt's paper is recommended as informative and stimulating. Results from studies of many workers are brought together to form an up-to-date picture, outstanding problems are clearly outlined, and promising leads to some solutions are suggested. Although in its bedrock geology this plateau region "ranks as one of the better known parts of the United States", Hunt's analysis reveals the Cenozoic history as a challenging frontier for geologic research.

CHESTER R. LONGWELL

REVIEWS

Risk and Gambling, the Study of Subjective Probability; by JOHN COHEN and MARK HANSEL. P. x, 153; 6 figs. New York, 1956 (Philosophical Library, $3.50).—Let it be stated at the outset that this little book by the two psychologists from the University of Manchester is not primarily a study of the lure and lore of gambling. Rather it is a report on the concepts of probability as they may be found among children and adolescents. The conclusions are based upon a number of estimations and guessing games conducted in school classes, and various emerging facts are quite interesting. For instance, young children 6-7 years of age tend to alternate their guesses regardless of successes or failures in previous attempts. The next age group, 8-10 years, alternates less but there is a preference for certain simple patterns in their guesses; only in the age group 12-14 years does there emerge some feeling of probability rules and reliance upon previous experiences. The material on the whole is small, and it would be of interest to see it expanded, also in connection with other types of experiments.

The modern theory of games and ecomnomic behavior assumes that individuals should be capable of making the most advantageous probability choices. However, the reviewer should be willing to venture the guess that, were the previous experiments to be applied to adults, they would on the whole show only the most rudimentary feelings for the laws of chance.

OYSTEIN ORE

Echt oder Synthetisch?; by K. F. CHUDOBA and E. J. GUBELIN. P. 156; 117 figs., 11 tables, 1 color plate. Stuttgart, 1956 (Ruhle-Diebener-Verlag K. G., DM 1850).—This concise and well organized book tells the reader a fascinating story, not only of how to distinguish between synthetic and natural gemstones, but also something of the reasons for the tests and something of the history of gem synthesis. The fact is, of course, that synthetic and natural gemstones are certainly identical in all essential physical and chemical properties; it is generally the imperfections of the formation of each that must be detected to determine its origin. Chudoba and Gübelin have here assembled

an authoritative and useful manual on such imperfections and their applications. The publisher is to be congratulated on the production of two such excellent, related books as this and F. W. Eppler's translation of B. W. Anderson's *Gem Testing* (Praktische Edelsteinprüfung, Stuttgart, 1955). Although there is some duplication, the two volumes complement one another and will find much use in the workshops of jewelers and of collectors of precious stones, as well as in the libraries of institutions and individuals interested in the study thereof. HORACE WINCHELL

Economic Geology Fiftieth Anniversary Volume, 1905-1955, Parts 1 and 2; ALAN M. BATEMAN, editor. Twenty-five papers plus foreword. Lancaster, Pennsylvania, 1955 (Economic Geology Publishing Co., $8.00).—The Fiftieth Anniversary Volume is a splendid tribute to the usefulness of the periodical whose name it bears and to the vision of its founders, now enshrined among the fathers of this branch of geology in America—Lindgren, Spurr and Kemp, Foster Bain, Marius Campbell, Ransome, George Oitis Smith, Ries, W. H. Weed, and H. V. Winchell, and especially its pioneer editor, J. D. Irving. Though appearing a bit later, W. S. Bayley, for so long its business manager, should be mentioned too in these lists of early supporters. The publication here reviewed encompasses in two generous volumes, of some 500 pages each, most of what is good from those earlier days of "Economic Geology" and much that is founded upon studies by these men and their contemporaries but much also that is new and that departs in valuable directions from the more classical tradition.

The opening chapter, with its well planned summary by the present editor, Alan M. Bateman, gives a picture of the history of achievements of the journal in significant trends in the various fields of applied geology, with special emphasis, as might have been expected, upon ore deposits. Among the contributors to the subsequent chapters are the followers and understudies of many of those who wrote for "Economic Geology" in its first two decades. Of special interest, perhaps, to those concerned with the genesis and discovery of ores are the chapters by Turneaure on ore provinces, by Donald E. White on the relations between hot springs and ore deposits, by Lovering on temperatures accompanying intrusions, by Schwartz on hydrothermal alteration as an ore guide, and the excellent and up-to-date review by Ingerson on geologic thermometry. But to mention some authors and subjects and omit others is, by implication, a discrimination that reflects the interest of the reviewer, rather than the merits of individual chapters.

The plan of the volume has been well conceived and is consistent with the purposes stressed in "Economic Geology" as these developed—a stress upon ore deposits but discriminating through briefer attention to kindred aspects of applied geology, such as engineering geology (with chapters by Robert Legget and Terzaghi), certain minor non-metals (clay, limestones, and pegmatites, as represented by chapters whose authors are respectively, Ralph E. Grim, D. L. Graf and J. E. Lamar, and R. H. Jahns), a section on the time aspects of petroleum accumulation by Levorsen, an excellent summary of the modern approach to ground-water problems (by J. G. Ferris and A. N. Sayre),

a chapter on the petrology of coal, now receiving so much attention, by C. E. Marshall, and several chapters having to do with special applications and developments in the general field—the synthesis of minerals, the uses of geophysical methods, paleoecological procedures applied to economic problems, and the like.

Little can be said in this limited space of the content of each of the 25 chapters or of the detailed contribution by each of the many authors. At best, as he reads, the critical scientist will become aware of certain sparks that can be seen as they flash from the anvils on which so much of the ore of the past fifty years is hammered into the necessarily closely wrought fabric of these two books. Perhaps the most uniformly useful contributions of the authors are of two types.

First, each contributor has been compelled, as it were, to change the depth of focus so common to those engaged in the application of geology to economic problems. Instead of seeking detailed facts about a single region or limited problem, the author of one of these chapters has been compelled to distill and redistill. This has given a certain pithy quality to the volume that becomes increasingly conspicuous as the reader traverses chapter after chapter. The views are, in a sense, broad rather than deep. They serve thus to compare what occupies the scholar of today in contrast with the founders of the journal and their contemporaries. With this goes a healthy skepticism, a tentative attitude of mind—the best foot forward of each writer, even though at other, less lucid moments and elsewhere he may have been guilty of a narrower view. Here the reader finds not the attitude, fortunately, of the protagonist and defender but rather that of the teacher or author of textbooks. A periodic review, dispassionate but learned, is badly needed and is happily supplied in most chapters; those on metallogenic zoning and on geologic thermometry are commendable examples.

A second invaluable contribution is the detailed and up-to-date nature of most of the bibliographies that follow the chapters. Each of these is a mine in itself, as quick references to those appended to the articles on ore structures and pegmatites will show. If a complaint is permissible, it is perhaps only that so few, relatively, of the works cited by certain authors are foreign; geology has by now lost much of its parochial aspect, and the end of the Second World War has made available the thoughts of Germans and Russians which are less frequently quoted than might be, even in this volume. The bibliographies in the chapters on ground water and on coal petrology are especially praiseworthy exceptions.

All in all this is a highy commendable effort; the authors are to be congratulated, and so is the research worker and teacher who finds here, ready to hand, a splendid supplement to his stock of textbooks on economic, applied geology. CHAS. H. BEHRE, JR.

Traité de Paléontologie, Tome V: Amphibiens—Reptiles—Oiseaux; by JEAN PIVETEAU and 12 others. P. 1114; 979 figs. 7 pls. Paris, 1955 (Masson et Cie, 12,800 francs.)—For more than 80 years the textbook of paleontology by Karl von Zittel has been a standard reference on the classification and dis-

tribution of fossils. Translated into various languages, and revised by a series of noted students, it has been the handbook for several generations of paleontologists. Inevitably such a compilation must become obsolete; new discoveries, changing views on major classification, and emphasis upon more recently developed branches of the science all gradually diminish the usefulness of such a treatise. For a quarter of a century a need has been felt for a more up-to-date general reference on all groups of fossils. Shortly following the close of World War II, work was commenced on the *Traité de Paléontologie* under the direction of Professor Jean Piveteau of the Sorbonne, in Paris. At about the same time, in the United States, preparation of a comprehensive *Treatise on Invertebrate Paleontology* was started under the leadership of Professor R. C. Moore of the University of Kansas. In view of the magnitude of these undertakings, and the inevitable duplication of effort in the treatment of the invertebrates, it seems unfortunate that paleontologists of the entire world did not combine efforts to produce a single treatise superior to any that a limited number of collaborators could compile. Be that as it may, the broad scope of the *Traité de Paléontologie*, which includes vertebrates and ancient man as well as invertebrates, should commend it to students in all countries.

Tome V of the *Traité*, comprising amphibians, reptiles with the exception of the synapsid or mammal-like branch, and birds, is actually the fourth part to appear. Tomes I to III which cover the invertebrates, appeared in 1952 in 1953; the fourth volume, on fishes, is in press; two other volumes, on mammals, and on fossil primates and early man, are in preparation. Mammal-like reptiles apparently are to be discussed along with mammals in Tome VI.

The present volume, by 13 collaborators, departs radically from the brief listing of fossil genera which forms so prominent a part of the Zittel handbook. Detailed abstracts are given of the most recent research on various groups. Where the compilers have preferred to depart from the usage of their sources, reasons therefor are explained in some detail. Morphologically well known genera are emphasized and described and other forms are compared with them. Following the classification of each group is an account of its geologic occurrence. Some sections include remarks on the associated faunas of important localities. In other chapters sections are devoted to special topics such as ecology, habits, distribution, soft anatomy, evolution, and extinction of particular vertebrate groups. Brief chapters summarize the representation of lower vertebrates in the art of primitive man.

The volume is notable for the attention paid to problems about which controversy exists at present. Arguments on both sides of conflicting theories on the origin of tetrapods, homology of dermal bones, relationships of the branchiosaurs, origin of birds, and many other disputed subjects are presented. On some of these the authors take definite positions, others are left open, but opposing views are fairly stated throughout.

A particularly important feature is the extensive and up-to-date bibliography following each chapter. Likewise useful are the historical résumés, not always complete but generally mentioning the earliest discoveries and most recent revisions, from which the interested reader would be led to more detailed accounts. Recent discoveries in Russia and Mongolia, not generally

known in America, are cited and incorporated into the discussion of labyrinthodonts and dinosaurs.

Inevitably the treatment by various authors is dissimilar. Detailed accounts of the osteology of recent crocodilians, frogs, and birds are not balanced by corresponding detail about salamanders, lizards, or snakes. This partially reflects the current state of knowledge of the anatomy of these existing Orders. The Gymnophiona are not mentioned at all which seems a curious omission even though they are scarcely known fossil.

These contributions have been skillfully linked together by Professor Piveteau. There is seldom any lack of continuity in passing from one section to another. A possible omission due to different authorship is the failure to compare *Askeptosaurus* with *Pleurosaurus*, two reptiles not too distantly related and with some similarities of skull construction. Some sections are more like abstracts of the best available works than true syntheses; this is inevitable in an encyclopedic treatment. A few sources quoted were not the latest and most authoritative; for example the reconstruction of *Camarasaurus* by Osborn and Mook, copied on p. 825, has been corrected subsequently by Gilmore.

The section on birds is disappointing. After an excellent summary of DeBeer's important recent study of *Archaeopteryx* and a review of the problem of the origin of birds, the fossil record of modernized birds is incompletely summarized and only partially interpreted. The outmoded separation of ratites and carinates is perpetuated although its artificiality is acknowledged. Major attention is paid to the ratites, but Macdowell's important study of their skulls is not mentioned, nor is the work of Archey on the moas cited. Many records of North American fossil birds are ommited from the summary, and there is no reference to Wetmore's check-list nor to the work of the Millers or Howard. This incompleteness contrasts markedly with the thorough treatment of the reptilian and amphibian orders.

This volume provides a complete and detailed account of the lower tetrapods, abundantly and well illustrated and clearly printed on good paper. Like the rest of the series, it is an extremely useful reference for the paleontologist and a reliable source of paleontological data for students of related sciences.

JOSEPH T. GREGORY

Publications Recently Received

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